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HYDROGEOLOGY OF A BURIED VALLEY AND ADJACENT AQUIFERS IN
SOUTHWEST POTTAWATOMIE COUNTY, KANSAS

by

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B. A., College of Saint Thomas, 1982

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A MASTERS THESIS

submitted in partial fulfillment of the

requirements for the degree

MASTER OF SCIENCE

Department of Geology

KANSAS STATE UNIVERSITY
Manhattan, Kansas

1988

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INTRODUCTION

Groundwater from the Paleozoic bedrock of northeast Kansas is often unsuitable for domestic and livestock use because it contains high concentrations of dissolved minerals. Almost ninety percent of the potable ground water in northeast Kansas is from Quaternary age aquifers (Frye and Walters, 1950).

Many Quaternary age aquifers, however, are inadequate sources due to low quantity or poor quality of water. Glacial till and outwash aquifers supply small and erratic amounts of water. In addition, water from glacial till aquifers is often contaminated by water from nearby bedrock, and may be unusable (Walters, 1953; 1954; Heath, 1984). Terrace and floodplain alluvium occur in the bottom of valleys, but they often do not occur in the areas where water is needed. The quality and amount of water from the terrace and floodplain alluvium vary greatly with season. During periods of low recharge the water quality becomes poor and the alluvium may not contain enough water to supply a municipality (Heath, 1984; Whittemore and others, 1982). In many areas, the most satisfactory sources of water are buried, pre-glacial valley aquifers because of their high yields and quality (Denne, 1982).

Buried, pre-glacial valleys are stream valleys that were filled with glacial sediments and that no longer have a stream flowing on their surface. The existence of buried, pre-glacial valleys has been known for many years, but the value of these deposits as aquifers has been recognized only recently (Denne, 1979).

The Kansas Geological Survey and the Department of Geology, Kansas State University, are conducting a joint investigation of the buried,

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pre-glacial valleys in northeast Kansas. The glaciated area of north-
east Kansas was divided into an eastern subregion that is being inves-
tigated by the Kansas Geological Survey, and a western subregion that
was being studied under the direction of Dr. Henry V. Beck, Department
of Geology, Kansas State University. The buried, pre-glacial valley in
southwest Pottawatomie County is being studied as a part of the west-
ern subregion.

Purpose of Investigation

The purpose of this investigation is to establish a data base for
the future study of an important aquifer. The differences in the
groundwater composition of the aquifers surrounding the buried, pre-
glacial valley are used to evaluate the possibility that the buried,
pre-glacial valley aquifer may be contaminated by water from other
aquifers. Finally, the seasonal variations of the composition of
groundwater from the buried, pre-glacial valley aquifer and the sur-
rounding aquifers are examined.

Location of Area

The area of investigation occurs in the southwest corner of Potta-
watomie County (Fig. 1), extending from the northern boundary of T. 9
S. south to the Kansas River and from two miles west of the western
boundary of R. 9 E. to the eastern boundary of R. 10 E..

Physiography and Land Use

This region contains parts of two physiographic sections of the
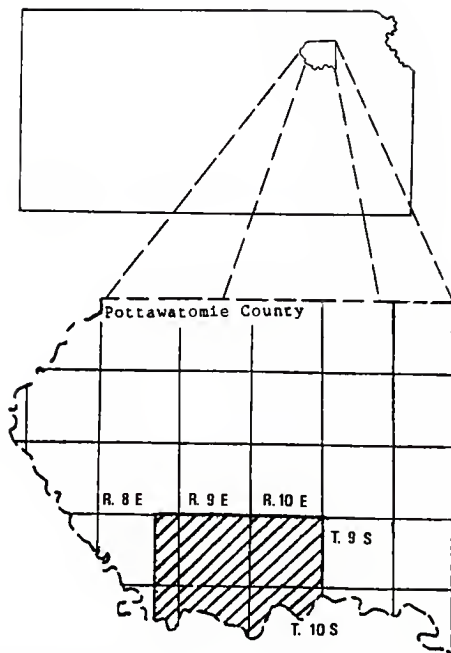
Figure 1.   Location of the area of investigation in Pottawatomie
           County, Kansas.

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Central Lowland Physiographic Province (Fig. 2). The southwestern
third of the area is within the Flint Hills Upland of the Osage Plains
Physiographic Section. The remaining portion of the area is within the
Attenuated Drift Border of the Dissected Till Plains Physiographic
Section (Schoewe, 1949). The boundary between the glaciated and non-
glaciated regions of the area coincides with the boundary between the
two physiographic sections (Scott and others, 1959).

The maximum elevation of the area is 1350 ft and the minimum ele-
vation is 970 ft. The change in elevation of the area is 400 ft and
the local relief is commonly between 150 and 250 ft. Tall grass prai-
rie covers most of the upland areas. Floodplain forests consisting of
oak, cottonwood and elm trees occur in the lower parts of watersheds
(Soil Conservation Service, 1980; 1975).

The temperature has a large range and changes rapidly. The average
winter temperature is 28 F (-2 C), and the average summer temperature
is 76 F (24 C). The average annual temperature is 52 F (16 C). The
average annual rainfall is 32 in.(81 cm). The winter and summer have
the least precipitation, and the late spring and the late fall have
the greatest precipitation. (National Oceanic and Atmospheric Admini-
stration, 1986).

The Kansas River is the principal stream and the southern border
of the area of investigation. The largest streams are Vermillion and
Rock Creeks (Fig. 2). Adams, Brush, Hopkins, and Elm Slough Creeks
empty into Rock Creek which flows into Vermillion Creek east of Louis-
ville. Vermillion, Elbo, Blackjack, and Sand Creeks are tributaries of
the Kansas River. The major towns are Wamego, St. George, and Louis-

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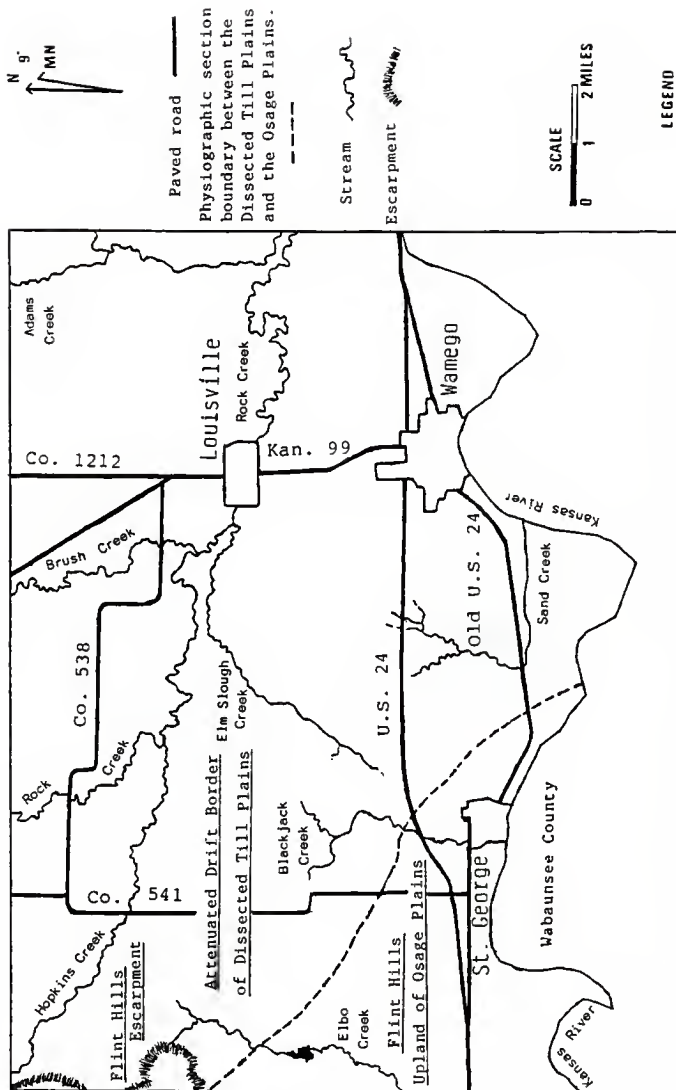
Figure 2. The physiography of the area of investigation. (After Scott and Others, 1959)
The area to the northeast is glaciated

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ville (Fig. 2). Two major highways cross the area: U.S. highway 24 and
Kansas highway 99. Paved County highways 1212, 538, 541 and old U.S.
24 and unpaved roads provide additional routes.

Agriculture is extensive in the stream valleys and the large por-
tions of the glaciated uplands that are covered by fine-grained sedi-
ments are extensively cultivated with diverse crops. The major crops
are sorghum, wheat, hay and corn. Irrigation is common in the Kansas
River Valley (Soils Conservation Service, 1975; 1980).

Livestock provide fifty percent of the revenue of the county. The
major type of livestock raised are cattle, but pigs and sheep are also
raised. Cattle and sheep are grazed in the upland areas which are un-
suited for cultivation. Pigs are raised where food can be cultivated
for them (Soils Conservation Service, 1975; 1980).

Previous Investigations

The structural geology of regions that contain Pottawatomie Coun-
ty have been investigated by Jewett (1951), Rieb (1954), Swett (1959)
and Chelikowsky (1972). The stratigraphy of Kansas sedimentary rocks
has been summarized by Moore and others (1951), Jewett (1963), and
Zeller (1968).

The Pleistocene stratigraphy and geology of Kansas was described
by Frye and Leonard (1949; 1951; 1952). The Pleistocene geology of
Kansas has been described by Hay (1893), Smyth (1898), Todd (1908),
Schoewe (1930), Frye and Walters (1950), and Aber (1985).

Several regional studies have been made of the groundwater re-
sources of northeast Kansas. Moore and others (1940) described the

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groundwater resources of Kansas. The Kansas Water Resources Board (1959) studied water resources in Kansas and the existing water problems of Kansas.

As part of a cooperative effort, the Kansas Geological Survey and the United States Geologic Survey studied the groundwater and geological resources of many counties in Kansas. The geology of Pottawatomie County was described by Scott and others (1959). Jewett (1941), Walters (1953; 1954) and Ward (1974) studied the groundwater resources and geology of several Counties bordering Pottawatomie County. The geology of the other counties bordering Pottawatomie County have been described by Mudge and Burton (1959), Mudge and others (1959) and Johnson (1967).

The geology and groundwater resources of the Kansas River have been extensively studied by Davis and Carlson (1952), Smith (1959), Beck (1959), Moulthrop (1960), Fader (1974) and O'Brien and others (1972).

Buried, pre-glacial valleys in Kansas have been described by Dreezen and Burchett (1971), Denne (1979), Chelikowsky (1976), Carl (1980), and Denne and others (1982). The buried, pre-glacial valley in southwest Pottawatomie County has been investigated by Hay (1893), Smyth (1898), Todd (1908; 1911; 1917), Mudge (1955), Scott and others (1959), Mudge and Burton (1959), and Snow (1963).

Whittemore and Switek (1977) and Switek (1977) analysed the surface water chemistry and trace elements in clays of a drainage in Pottawatomie County. The cities of St. George and Wamego, the Kansas Department of Health and Environment, and Pottawatomie County Rural

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Water District Number One have water wells in the area of investigation. The Kansas Department of Health and Environment has data on the water quality of the wells listed above (appendix I).

Paleozoic Geology of the Area

Structure--The Nemaha Anticline trends across the area of investigation in a south-southwesterly direction and exposes Pennsylvanian Rocks 15 miles farther to the west than expected (Fig. 3; Scott and others, 1959; Mudge and Burton, 1959). The strata on the eastern flank of the Nemaha Anticline have an eastward dip of nearly 3 degrees over a distance of four miles. The strata on the western flank dip 3 degrees westward over a distance of four miles (Rieb, 1954).

Two sets of joints occur in the Paleozoic rocks in the area. One set of joints trends northeast to southwest; the second set of joints trends northwest to southeast (Chelikowsky, 1972). The angle between the joint sets is almost 90 degrees. The joints form zones of weakness that water flows through and that some streams form along (Chelikowsky, 1976).

Paleozoic History--Shallow marine sediments were deposited on crystalline bedrock in the area during the early Paleozoic. Uplifts formed during the Late Cambrian and Late Ordovician, and the sedimentary rocks over the axis of the uplifts were removed by erosion (Ratcliff, 1957; Lee, 1943).

The Paleozoic bedrock of the area formed during the Permian and Pennsylvanian Periods when shallow inland seas again invaded the mid-

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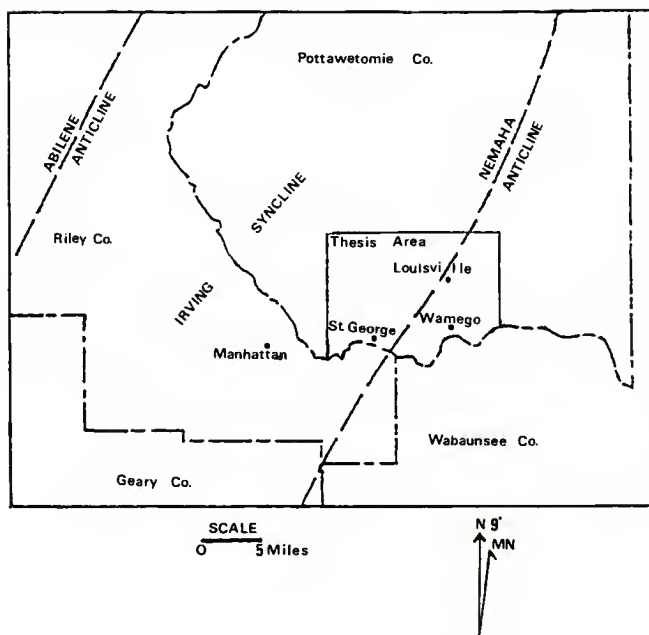
Figure 3. The structural geology of Pottawatomie County, Kansas.
(Modified from Chelikowsky,1976)

continent area of North America and the deposition of shallow marine carbonates and terrigenous sediments alternated with episodes of erosion. The Nemaha Anticline formed during the Late Mississippian when older faults were reactivated. (Lee, 1943; Koons, 1956; Mudge and Burton, 1959). The inland seas withdrew at the end of the Paleozoic Era (Scott and others, 1959; Mudge and Burton, 1959).

Paleozoic Stratigraphy--The Paleozoic bedrock of the area consists of the alternating shales and limestones of the Wabaunsee Group of Upper Pennsylvanian Age and the Admire, Council Grove, and Chase Groups of Lower Permian Age (Fig. 4).

The Wabaunsee Group is made up predominantly of shale and thin limestone beds (Fig. 5). Many of the shales of the Wabaunsee Group are sandy (Scott and others, 1959; Beck, 1959), and channel sandstones may occur in several beds of shale in the Wabaunsee Group (Mudge, 1955). A widespread blanket sand in the top of the Willard shale outcrops between St. George and Wamego (Voss, 1972). The sandstone in the Willard Shale, and the Elmont, Burlingame, Dover, Tarkio, and Brownville Limestones produce small to moderate amounts of water (Voss, 1972; Beck, 1959; Smith, 1959).

Channel sandstones are especially common in the lower shales of the Admire Group where they cut into the Wabaunsee Group, and many units of shale contain basal sands (Fig. 5; Moore and others, 1951; Scott and others, 1959; Beck, 1959). The only unit in the Admire Group to be used as an aquifer is the Falls City Limestone (Beck, 1959; Smith, 1959).

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LEGEND

Quaternary System
- Floodplain Alluvium
- Terrace Deposits
- Bignell
- Peoria and Loveland Loess
- Qgl — Fluvial and Lacustrine deposits
- Qt — Till
- Qg — Pre-Kansan Gravel

Permian System
- Pc — Chase Group
- Council Grove Group
- Pa — Admire Group

Pennsylvanian System
- Wabaunsee Group

SCALE
0   1   2 MILES
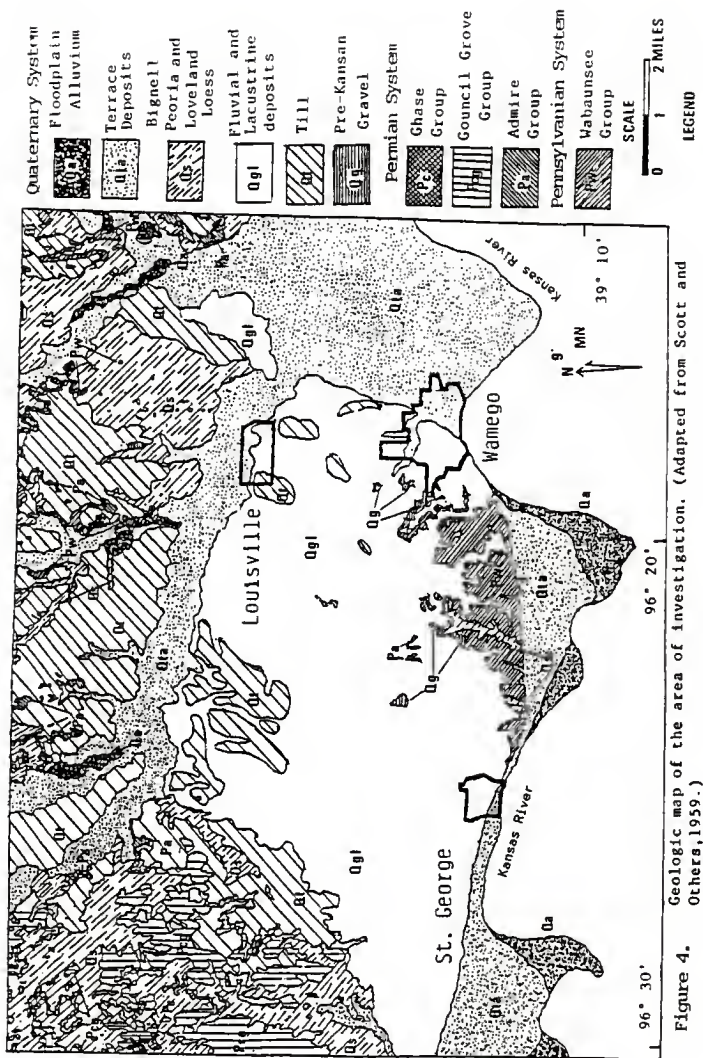
Figure 4.   Geologic map of the area of investigation. (Adapted from Scott and Others, 1959.)

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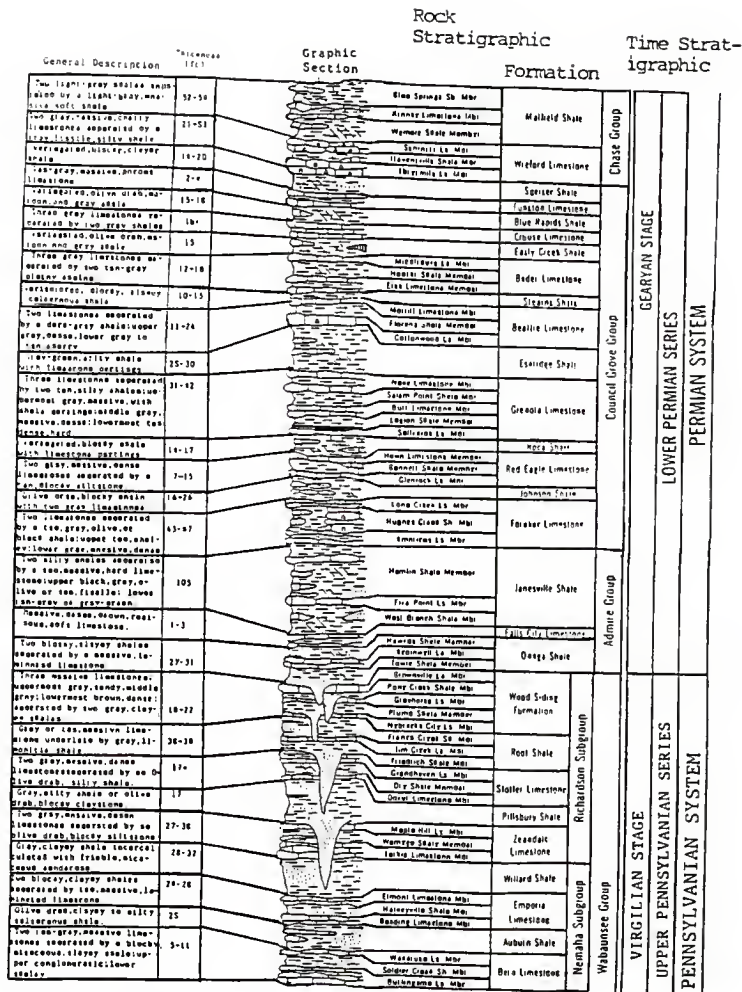
Figure 5. Paleozoic stratigraphic section of the area of investigation. (Modified from Zeller, 1968; descriptions from Beck, 1959 and Scott and others, 1959)

The units of the Council Grove Group consist of two limestones separated by a thin intervening shale alternating with thick, varicolored shales (Fig. 5; Zeller, 1968). The shales of the Council Grove group are capable of being aquifers because the weathered zones of the calcareous shales are capable of supplying small amounts of water to a well (Beck, 1959). The most frequent aquifers of the Council Grove Group are the Cottonwood, Burr, Neva, Howe, and Long Creek limestones (Beck, 1959; Smith, 1959).

The Wreford Limestone and the Matfield Shale of the Chase Group are in the area of investigation, but they are not aquifers (Fig. 5; Scott and others, 1959). The Wreford Limestone contains two light gray, extremely cherty limestones separated by a gray calcareous shale. The limestone members of the Wreford Limestone are the bedrock sources of the pre-Kansan gravels (Scott and others, 1959).

Cenozoic History

Northeast Kansas was glaciated by glaciers of Nebraskan and Kansan age (Frye and Walters, 1950; Frye and Leonard, 1952). Loess was deposited on the flat-lying hilltops of the area during the Kansan, Illinoian, and Wisconsinan Glaciations (Scott and others, 1959). Alluvium was deposited in the major stream valleys during the Nebraskan, Kansan, Illinoian, and Wisconsinan ages. Glacial Till was deposited during the Nebraskan and Kansan ages (Frye and Leonard, 1952; Scott and others, 1959).

The pre-glacial Kansas River was the main stream in the area before the Kansan glaciation, and is believed to have formed the bu-

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ried, pre-glacial valley in southwest Pottawatomie County (Davis, 1951; Mudge, 1955; Scott and others, 1959). The pre-glacial Kansas River drained the western edge of the Kansan glacier when glacial meltwater reached the headwaters of its northern tributaries (Todd, 1908; 1917; Scott and others, 1959).

The Kansan glacier entered the study area from the northeast and extended as far west as the Flint Hills (Fig. 6). The front of the Kansas Glacier extended southeastward from the Flint Hills into Wabaunsee County and dammed the pre-glacial Kansas River to the east of St. George (Fig. 6; Scott and others, 1959; Mudge, 1955; Mudge and Burton, 1959). Meltwater and outwash from the Kansan glacier filled the pre-glacial Kansas River valley to the west of St. George and formed Lake Kaw (Hay, 1893; Smyth, 1898; Todd, 1918; Mudge, 1955; Scott and others 1959). Glacial Lake Kaw drained east through ice marginal streams (Todd, 1908; 1911; 1917; Mudge, 1955; Scott and others, 1959; Aber, 1985).

The Kansan glacier retreated to the northeast and formed a stable front roughly parallel to the present day Rock Creek. Meltwater from the southern margin of the Kansan glacier flowed to a valley between St. George and Wamego because outwash dammed the pre-glacial, Kansas River Valley. East of St. George the meltwater followed a newly opened valley to the northeast and joined the pre-glacial, Kansas River valley east of Wamego (Scott and others, 1959). Glacial meltwater from the western margin of the Kansan glacier flowed in the present Kansas River valley and joined the meltwater flowing to the northeast (Scott and others, 1959).

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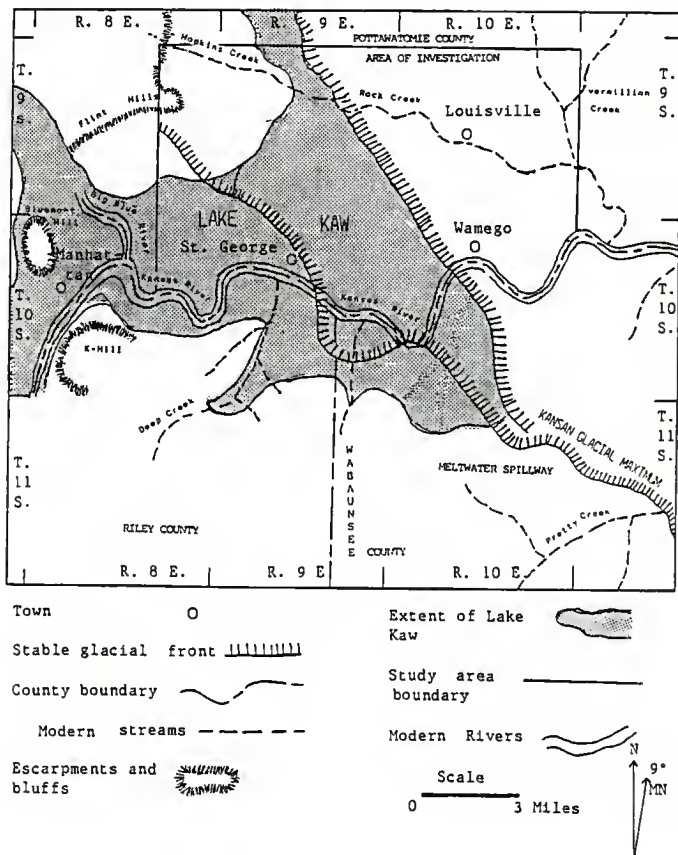
Figure 6. Modern drainage, stable glacial fronts, and the extent of Lake Kaw. (Modified after Todd, 1918; Mudge, 1955 and Scott and others, 1959.)

The Kansan glacier made a small advance shortly before it re-
treated that left glacial till deposited on top of the buried valley
sediments. The glacial till dammed the drainage between St. George
and Wamego, and the meltwater followed the present course of Rock
Creek (Scott and others, 1959).

Quaternary Stratigraphy

Investigations of the Quaternary stratigraphy of Pottawatomie
County predate the investigations that have led Quaternary geologists
to conclude that at least 7 glaciations have occurred during the last
2.5 million years (Boellstorff, 1978). Classic Quaternary time strati-
graphic terms are used in this report because there are no commonly
accepted time stratigraphic terms corresponding to the new concepts of
glaciation to replace them.

The Quaternary sediments are unconsolidated and discontinously co-
ver the Paleozoic bedrock. The oldest Quaternary sediments are chert
gravels that are pre-Kansan in age. The youngest Quaternary sediments
are alluvial deposits in the stream and river valleys. Most of the
Quaternary deposits of the area were deposited during the Kansan stage
of the Pleistocene Epoch (Scott and others, 1959).

Pre-Kansan Gravels--Pre-Kansan gravels occur in alluvial terraces
that are found from 80 to 120 ft (27-40 m) above the present stream
valleys (Fig. 7; Frye and Leonard, 1949; 1952; Davis, 1951; Davis and
Carlson, 1952; Mudge, 1955). Pre-Kansan gravel deposits are also be-
lieved to occur in the bottom of the buried, pre-glacial valley (Beck,

Rock-stratigraphic
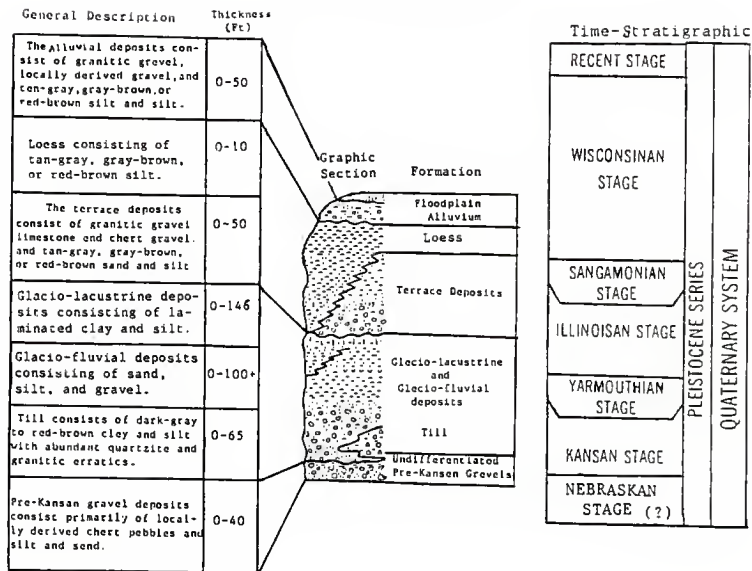
Figure 7. Quaternary stratigraphic section of the area of investigation. (Modified after Beck, 1959; Scott and others, 1959 and Zeller, 1968).

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Personnel Communication, 1985).

Pre-Kansan gravels are chert gravels that are believed to have been deposited before the Kansan Glaciation (Fig. 7). The ages of the gravels are unknown, but estimates range from Triassic to early pre-Kansan in age (Davis, 1951; Beck, 1959). Of course, the ages of pre-Kansan gravels at different localities may be different.

Unlike gravel deposits formed during and after the Kansan glacial stage, the pre-Kansan gravels contain very few erratics or sediment derived from the Ogalalla Formation (Davis, 1951; Frye and Leonard, 1952; Beck, 1959). The pre-Kansan gravels consist mainly of locally derived chert, but some deposits contain large amounts of locally derived limestone. The gravels are subangular to subrounded and have a maximum grain size of 3 ins (10 cm) in diameter (Scott and others, 1959). The gravels are cemented by slightly sandy, reddish brown clay (Scott and others, 1959; Beck, 1959; Mudge, 1955). A few erratics may occur in the top of some pre-Kansan gravel deposits indicating that the Kansan ice sheet may have overridden the gravels or that the gravels are early pre-Kansan in age (Scott and others, 1959).

The pre-Kansan gravels are exposed on the north side of the Kansas River Valley in high divides between Wamego and St. George (Fig. 4; Scott and others, 1959). The pre-Kansan gravels range from a feather-edge to 10 ft (3 m) and average 5 ft (2 m) in thickness (Scott and others, 1959; Beck, 1959). The gravels are overlain by glacial till and glacial outwash and underlain by Paleozoic bedrock. The pre-Kansan gravels exposed between Wamego and St. George are believed to have been deposited by tributaries of the pre-glacial, Kansas River because

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they cannot be traced laterally any great distance (Mudge, 1955; Beck, 1959).

Glacio-fluvial and Glacio-lacustrine Deposits--A combination of in-
terbedded glacial lake sediments and glacial outwash of Kansan age
have filled the buried, pre-glacial valley in the area, and form a li-
near, silty-sand deposit flanked by finer-grained deposits (Snow,
1963; Mudge, 1955; Scott and others, 1959).

The glacio-fluvial and the glacio-lacustrine deposits are mapped
as a single geologic unit (Fig. 4; Fig. 7). The glacio-fluvial depo-
sits probably represent both the fluvial sediments deposited in front
of the glacier as it advanced and the outwash deposited when the gla-
cier retreated (Frye and Leonard, 1959). The glacio-lacustrine sedi-
ments were deposited in Lake Kaw and ice marginal lakes (Todd, 1908;
1911; 1917; Scott and others, 1959; Snow, 1963).

The grain size of the glacio-fluvial and glacio-lacustrine depo-
sits exposed on the surface of the buried, pre-glacial valley are very
similar (Snow, 1963). The glacio-fluvial deposits consist of strati-
fied sand, silt, and clay. The glacio-lacustrine sediments consist of
laminated clays and silt (Scott and others, 1959; Snow, 1963). The
sediments are composed predominantly of quartz with feldspar as a com-
mon constituent (Snow, 1963). The deposits may be iron stained and may
be locally cemented by calcium carbonate (Scott and others, 1959).

The glacio-fluvial and glacio-lacustrine deposits filling the bu-
ried, pre-glacial valley may be as much as 200 ft (67 m) thick (Mudge,
1955; Scott and others, 1959). Glacio-fluvial and glacio-lacustrine

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deposits occur east of Louisville and may represent a part of the buried, pre-glacial valley sediments that were separated from the main body of the sediments by the formation of Rock Creek (Scott and others, 1959).

Kansas Till--The Kansan Till consists of unstratified sediments that were deposited directly by the Kansan glacier as it retreated from the area (Fig. 7; Frye and Walters, 1950; Frye and Leonard, 1949; 1952).

The Kansan Till consists of unconsolidated deposits of sand, gravel, and boulders within a clayey to silty matrix (Frye and Walters, 1950; Frye and Leonard, 1952). The matrix was derived from the mechanical and chemical weathering of the shales and limestones of northeast Kansas (Scott and others, 1959; Frye and Leonard, 1952). The gravel and boulders are made up of locally derived limestone and chert, and erratics of granite, greenstone, and quartzite (Scott and others, 1959). The granite and greenstone erode easily and are more common in unoxidized till than they are in oxidized till (Beck, 1959). Vertical fractures in the till allow it to be undercut by running water, and it is easily eroded (Scott and others, 1959; Frye and Walters, 1950).

A thin, discontinuous layer of Kansan Till occurs in the northeastern two-thirds of the area (Fig. 4). The Kansan Till usually occurs on hills (Scott and others, 1959; Beck, 1959). The Kansan Till may be up to 40 ft (13 m) thick, and it disconformably overlies many of the older deposits (Beck, 1959). Glacial outwash deposits of clay, silt, sand, and gravel are often interstratified with the till.

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Loess--Loess is found mantling the tops of the highlands in the study
area, and it may be up to 20 ft (6.6 m) thick (Fig. 4; Beck, 1959;
Scott and others, 1959, Smith, 1959). It either was not deposited, or
it was removed by erosion in the rest of the area. Beck (1959) con-
sidered the loess in the area to be made up of the Loveland and Peoria
Formations, but he did not differentiate between the two formations
(Fig. 7).

In Pottawatomie County, the loess consists mostly of tan-gray,
gray-brown, or tan-brown, friable, clayey silt (Scott and others,
1959; Beck, 1959). In some places the loess is cemented by calcium
carbonate. The loess outcrops as vertical walls and becomes blocky
when it is weathered (Frye and Walters, 1950; Beck, 1959, Scott and
others, 1959).

The geologic map of the area includes many other types of deposits
mapped with the loess (Fig. 4; Scott and others, 1959). Any Quaternary
deposit that was too small to be mapped at the scale shown, and which
was adjacent to loess, was included with the loess on the map (Scott
and others, 1959).

Terrace, Floodplain, and Upland Stream Alluvium--Terrace and flood-
plain alluvium were deposited in stream valleys (Fig. 7; Beck, 1959;
Zeller, 1968). The terrace alluvium consists of the stream sediments
that are higher than the floodplain of the stream. Recent alluvium
consists of the sediments that make up the floodplain of the stream
(Zeller, 1968; Scott and others, 1959). Both the terrace and flood-
plain alluvium consist of reworked sediment from further upstream and

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from local tributaries. The major difference is that the terrace al-
luvium has a soil that is thicker than that of the floodplain of the
same stream (Scott and others, 1959; Soil Conservation Service, 1975;
1980).

The alluvial deposits of the major streams of the area are similar
to each other in grain size (Scott and others, 1959). The Kansas River
alluvium is composed of light gray to tan silty-sand that grades down-
wards into coarse sand and gravel (Moulthrop, 1960; Smith, 1959; Beck,
1959). The alluvium of the upland streams is made up of light gray to
tan brown, sandy-silt that grades downward into coarse sand and gravel
(Davis and Carlson, 1952; Scott and others, 1959).

The alluvial deposits of the major streams, however, differ in
their mineralogy. The sediments in the Kansas River alluvium are from
upstream and the gravel consists of quartz, feldspar, chert, lime-
stone, and granitic rock fragments that are subrounded to subangular
(Davis, 1951; Davis and Carlson, 1952; Moulthrop, 1960; Scott and
others, 1959; Beck, 1959). Most of the chert and limestone fragments
of the Kansas River alluvium are concentrated near the mouths of the
local tributary streams (Moulthrop, 1960). The alluvium of the upland
streams is locally derived, and it consists of limestone, chert, and
gravel from glacial till (Scott and others, 1959; Davis and Carlson,
1952). The limestone and chert fragments of the upland stream alluvium
are subangular to jagged (Davis and Carlson, 1952; Switek, 1977).

The thickness of the Kansas River alluvium is much greater than
the thickness of the upland stream alluvium. The alluvium of the
Kansas River ranges in thickness from a featheredge to 90 ft (30 m)

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and averages 40 to 60 ft (13-20 m; Beck, 1959; Davis and Carlson, 1952; Fader, 1974). The thickness of the upland stream alluvium ranges from a feather edge to 40 ft (13 m; Scott and others, 1959; Frye and Walters, 1950).

METHODS OF INVESTIGATION

Well Location and Numbering System

The well numbers used in this investigation are the locations of the wells according to the General Land Office system of land description. The parts of the well numbers are given in the following order: township, range, section, quarter section, quarter-quarter section, and quarter-quarter-quarter section (ten acre tract). The quarter sections, quarter-quarter sections, and ten acre tracts are designated a, b, c, or d in a counter-clockwise direction, beginning in the northeast corner with "a" . The well shown in figure eight would be numbered 9 10 7acd.

Water samples are numbered using the well number followed by a number designating the order in which the samples were collected from the well, starting with number one and proceeding upward numerically.

Maps and Stratigraphic Correlation Diagrams

Water well records submitted to the Kansas Department of Health and Enviorment by water well drillers were used to obtain Geologic information. Information from water well records and chemical analyses was contoured using the surface II computer mapping package (Sampson, 1975). Stratigraphic correlation diagrams of the buried valley were constructed from information obtained from the water well logs.

The base maps used in the study were the 7.5 minute series quadrangle maps compiled by the United States Geological Survey. The Louisville, St. George, Wamego, and Flush 7.5 minute quadrangle maps were used to establish well locations and to establish well elevations

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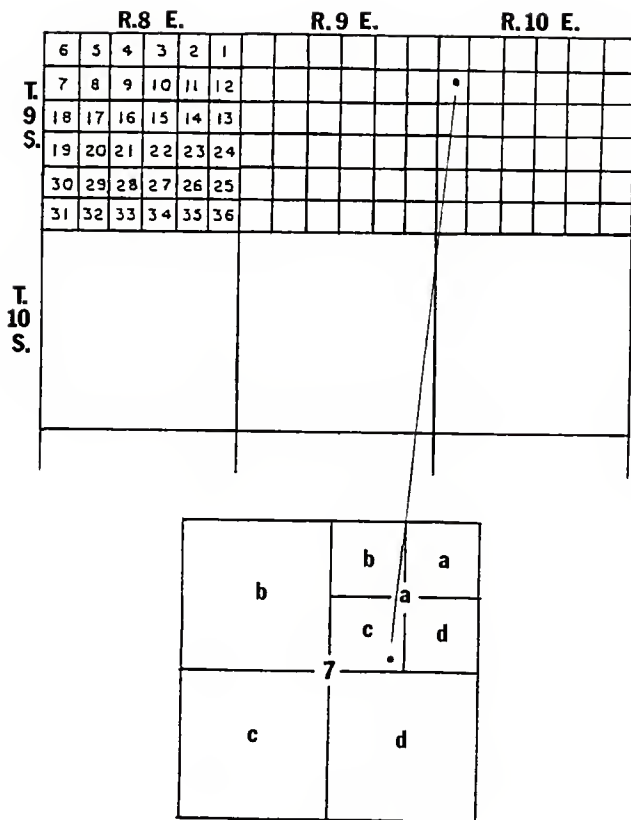
Figure 8. The location of a well using the General Land Office system of land description (9 10 7acd).

to the nearest five feet. When possible, the well site was visited to ensure that it was correctly sited (Appendix II).

Sampling Strategy

Four series of water samples were collected during a one year period. The total number of water samples taken were 70 (Appendix III). The first sample series of thirty water samples was collected during the period from July 23 to August 8, 1985. Samples were initially selected that represented the geographic range and aquifer type in order to examine the variation of the composition of the waters. The desired density of the sampled wells was one well every two miles for the buried valley sediments and one well every four miles for the other hydrologic units, which was not always achieved because of a lack of water wells in some areas. The final locations of the domestic wells for groundwater sampling were determined by their accessibility, and by the willingness of the well owner to allow the well to be sampled (Fig. 9).

The second and third sample series consisted of sixteen water samples collected from selected sites to provide data to examine possible variation in water composition with time (Appendix III). These wells were either representative of the aquifer types or contained high concentrations of dissolved solids. A large number of the sampled wells were from the buried valley aquifer to ensure that adequate data for this aquifer would be available. The second sample series was collected during the period from November 7 to November 15, 1985, and the third sample series was collected during the period from March 4 to

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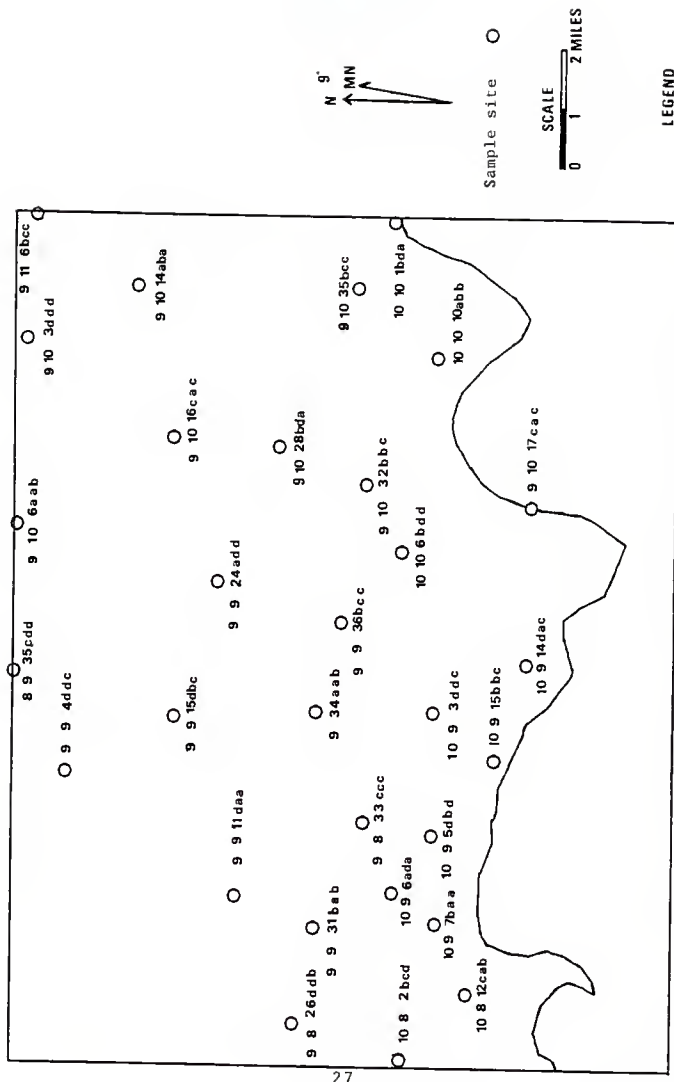
Figure 9. The location of water wells sampled during the investigation.

9 11 6bcc
9 10 14aba
9 10 3ddd
9 10 35bcc
10 10 1bda
9 10 16cac
10 10 10abb
10 10 4ddc
9 10 6aab
9 10 28bda
9 9 24add
9 10 32bbc
9 10 17cac
8 9 35cdd
9 9 36bcc
10 10 6bdd
9 9 4ddc
9 9 15dbc
9 9 34aab
10 9 3ddc
10 9 14dac
9 9 11daa
10 9 15bbc
9 8 33ccc
10 9 6ada
10 9 5dbd
9 9 31bab
10 9 7baa
9 8 26ddb
10 8 12cab
10 8 2bcd

N 9°
MN

Sample site   O

LEGEND

27

March 11, 1986.

The waters that showed the greatest chemical variation during the first three sampling series were sampled for the fourth sample series. Eight of the sixteen wells were sampled during the fourth sample series. The fourth sample series was collected during the period from June 15 to June 18, 1986 (Appendix III).

Field Procedures

The water samples were taken according to the procedures outlined by Whittemore (1983) and Switek (1977).

The sample bottles and other containers used for the collection of the water samples were washed with soapy water and rinsed with tap water and distilled water. The bottles and containers were filled with 1 normal nitric acid and allowed to soak for 24 hrs. The containers were rinsed with distilled water and given a final rinse with distilled and deionized water.

The water samples were collected from the water spigot nearest the well head. Each well was thoroughly flushed by pumping, and the pressure tanks were bypassed or drained before the sample was collected. Two 500 ml samples of water were collected from each well in 500 ml polyethylene bottles on each sampling occasion. Each sample bottle was rinsed three times with the sample water and rinsed once with filtered sample water before a sample was collected in the bottle.

The sample water was filtered through a 0.45 nm Gelman plastic membrane filter using a one liter pressure flask. The pressure flask

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was pressurized with a bicycle pump. One sample bottle was completely filled with water, and the second bottle was filled with 500 ml of water that had been immediately acidified with one ml of 6N HCl acid. The sample bottles were closed tightly and were marked with the sample treatment, and the name of the well owner. The water samples were immediately placed in an ice chest and then placed in a refrigerator to preserve the samples.

The temperature of the water from the well taken to the nearest tenth of a degree centigrade was obtained from a three gallon sample. The three gallon sample was used to minimize temperature change while the thermometer was in the container. The temperature was checked periodically during the flushing of the well until the temperature of the water from the well remained constant. The constant temperature indicates that the well pipe has been throughly flushed.

The pH of the water was measured at the well site. Two pH-meters were used to measure the field pH. The pH-meter used for the first, second, and third series was a Sargent-Welch Model PBX pH/mV meter with a Fischer Accu-Phast combination silver-silver chloride reference electrode. The pH-meter used for the fourth series of samples was a Cole-Parmer Model 5985-80 Digi-Sense pH-meter with a Cole-Parmer Model 5992-20 combination Electrode. The meters were calibrated using pH 4.0 and 7.0 buffers. The buffers partially overlap the expected range of the sample waters, and were available for use with the pH-meters. The pH was checked three times and the average of the two closest values has been given as the pH of the samples.

The specific conductivity was measured three times for each sam-

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ple, and the average of the two closest readings, corrected to 25 C, has been given as the specific conductivity of the sample. The specific conductivity of the water was obtained at the well site for the first, second, and third sample series. The specific conductivity meter used was a Lab-Line model Mark IV electro Mho-Meter with a Lab-Line model CD/10/D dipping epoxy conductivity cell with a cell constant of 1.03.

During the fourth sample series, the Lab-Line Electro Mho-Meter was not working. The specific conductivity of the fourth series of water samples was tested by collecting unfiltered, unacidified samples of water in polyethylene bottles, icing the bottles, and transported them to the Kansas State University Department of Agronomy where the specific conductivity of the water was determined using a Sargent-Welch Model 31 Conductivity Bridge with a cell constant of 1.0.

<center>Laboratory Procedure</center>

The major ions and selected trace ions were analyzed in the groundwaters. Detailed procedures of the analyses of each ion are given in Appendix IV, and types of analyses are summarized in Table 1. The acidified samples were analyzed for calcium, sodium, potassium, magnesium, strontium, iron, and manganese using a Perkin-Elmer 305B Atomic Absorption Spectrophotometer.

The unacidified samples were analyzed for chloride, sulfate, bicarbonate, and nitrate. The results of the nitrate analyses for sample series one are not given because the analyses were done incorrectly. The bicarbonate analysis was made within 24 hours of the sample col-

<center>30</center>

Table 1. The laboratory methods and instruments used in the analyses.

Dissolved Constituent	Laboratory Method	Instrument	Electrode
Sodium Magnesium Strontium Iron Manganese	Atomic Absorption Spectrophotometer	Perkin-Elmer 305B Atomic Absorption Spectrophotometer	
Calcium Potassium	Flame Emission Spectrophotometry	Perkin-Elmer 305B Atomic Absorption Spectrophotometry	
Chloride	Argentometric Titration		
Bicarbonate	Electrometric Titration	Corning Model 610 pH/mV meter	Fischer Accu-phast Combination Electrode
Sulfate	Turbidimetry	Coleman Model 14 Universal Spectrophotometer	
Nitrate	Nitrate Specific Ion Electrode	Corning Model 610 pH/mV meter	Orion Model 92-07 Nitrate Ion Electrode
			Beckman Calomel Reference Electrode

lection, and the nitrate analysis was made within three days to a week
of the sample collection.

Accuracy of Water Analyses

The accuracy of the water analyses was checked by two methods.
The first method compared the results of some of the analyses to the
results of analyses of the same samples by other laboratories (Appen-
dix V). Waters of the third series of samples were analyzed for cal-
cium, magnesium, sodium, and potassium by the campus emission spectro-
scopy laboratory (CESL) of Kansas State University. Three of the wa-
ters of the third series of samples were analyzed for the major cat-
ions and anions by D.O. Whittemore of the Kansas Geological Survey.

The comparison of two sets of analytical data for the same sam-
ples is accomplished by performing a t-test on the differences between
the two sets of data for the same set of samples (Harris, 1982). The
average difference and the standard deviation of the differences of
the two sets of analyses for a particular ion are needed for a t-test
and are found using the equations (Harris, 1982),

$$\overline{d} = \frac{\Sigma\, di}{n} \tag{1}$$

$$Sd = \sqrt{\frac{\Sigma\, (di - \overline{d})^2}{(n - 1)}} \tag{2}$$

where, n = number of pairs of samples

$\overline{d}$ = average observed difference

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di = observed difference for individual samples

Sd = standard deviation of the observed differences

The t value is found by comparing the standard deviation of the
data to the mean of the data using the equation (Harris, 1982),

$$t = \frac{\overline{d}\sqrt{n}}{Sd} \tag{3}$$

Using a table of t, the calculated value of t must be compared to
the values of t located in the row labelled with a number equal to the
number of pairs of samples analysed. The calculated t value must be
compared to the values of t in the table until it is found to fall be-
tween the t values of two neighboring columns. The probability that the
two analyses are different will fall between one minus the value at
the top of the left hand column and one minus the value at the top of
the right hand column. A probability greater than 95 percent is consi-
dered to be significant by statisticians (Harris, 1982).

A t-test comparing the analyses by the KGS and the analyses for
this study indicates that the analyses for Ca, Mg, and Cl differ from
each other significantly (Table 2). A t-test comparing the analyses by
the CESL and the analyses for this study indicates that the analyses
for Na and K are significantly different. The values of chloride in
the samples of this study were small and were analysed to the nearest
part per million (ppm). The chloride values given by the KGS are given
to the nearest tenth of a ppm. The rounding off of the chloride values
of this study may have significantly changed the chloride values of

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Table 2. The results of a t-test of the differences between the analyses conducted for the investigation and those by the Kansas State University CESL.

Ion	Average Difference	Standard Deviation of Differences	Calculated t Value	Number of Samples	Chance of Significant Differences
Calcium	-1.63	4.35	1.49	16	80 to 90 Percent
Magnesium	-.331	1.85	.717	16	50 to 60 Percent
Sodium	-4.25	9.38	1.81	16	90 to 95 Percent
Potassium	-.413	.378	4.37	16	99 + Percent

The results of a t-test of the differences between the analyses conducted for the investigation and those by the Kansas Geological Survey.

Ion	Average Difference	Standard Deviation of Differences	Calculated t Value	Number of Samples	Chance of Significant Differences
Calcium	-3.33	1.53	3.78	3	95 to 98 Percent
Magnesium	-.733	.351	3.62	3	95 to 98 Percent
Sodium	.400	.693	1.00	3	60 to 70 Percent
Potassium	-.100	.529	.327	3	20 to 30 Percent
Bicarbonate	-19.3	15.14	2.21	3	80 to 90 Percent
Sulfate	-1.77	2.78	1.10	3	60 to 70 Percent
Chloride	-.700	.265	4.58	3	98 to 99 Percent
Nitrate	-.233	2.55	.158	3	10 to 20 Percent

the two studies.

The great differences between the Ca, Na, K, and Mg values indicated by the t-tests may be due to many causes (Table 2). A likely cause of the differences is the great difficulty of analysing these particular ions. Another possible cause of the differences is the difference in analytical instruments used. The KGS and CESL both used newly acquired inductively coupled plasma emission spectrophotometers to analyse the samples, but the analyses for this study were performed using a much less sensitive atomic absorption spectrophotometer. Interestingly, the Ca, Na and K values of this study are between the Ca, Na, And K values of the KGS and CESL (Appendices V).

A second method of validating the accuracy of analyses containing more than 5.0 milligram equivalents per liter dissolved solids is by comparing the equivalents per liter of the anions to the equivalents per liter of the cations (Hem, 1970; 1985). The equivalents per liter of a particular ion is calculated by using the equation,

$$E = \frac{\dfrac{C}{F}}{I} \tag{4}$$

where E = equivalents per liter of the ion.

C = concentration of the ion in milligrams per liter.

F = formula weight of the ion.

I = charge of the ion.

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The sum of the equivalents of the anions should be compared to the sum of the equivalents of the cations according to the equation,

$$P = \frac{(C-A) \times 100}{(C + A)} \qquad (5)$$

where, P = percent difference between the anions and the cations

C = sum of the cations

A = sum of the anions

The percent difference of a perfect analysis would be 0. The percent difference of an accurate analysis is considered to be 5 percent or less (Hem, 1970).

The percent differences of the analyses of the second, third, and fourth series of samples were calculated to check the accuracy of the analyses (Table 3; Appendix VI). The percent differences for sample series one were not calculated because of the incomplete analyses. The average percent differences of the analyses of each sample series are all less than 5.0 percent.

The average percent differences of the analyses of the sample series are negative values, indicating that the results of the anion analyses were larger than the results of the cation analyses (Table 3). This could be due to the interference of bicarbonate and chloride in the nitrate analyses. Alternately it could be due, in part, to changes in the bicarbonate in the water samples during the 24 hours between the sample collection and the analyses for bicarbonate.

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Table 3. The mean of the percent difference and the mean
of the absolute value of the percent difference of the
water samples from sample series two, three, and four.

Sample Series Two

number of samples	mean of the abso-lute value of the percent differences	mean of the percent differences
16	3.0 percent	-2.5 percent

Sample Series Three

number of samples	mean of the abso-lute value of the percent differences	mean of the percent differences
16	6.1 percent	-1.5 percent

Sample Series Four

number of samples	mean of the abso-lute value of the percent differences	mean of the percent differences
8	5.4 percent	-3.0 percent

Sample Series Two, Three, and Four

number of samples	mean of the abso-lute value of the percent differences	mean of the percent differences
40	4.7 percent	-2.2 percent

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RESULTS

Geology

The extent of the surficial deposits of the buried, pre-glacial valley are well known, but the topography of the bedrock surface underlying the deposits has never been examined. Speculations as to the topography of the buried, pre-glacial valley have been made: Todd (1917) believed that it would run in an east-west direction somewhere between one and three miles north of St. George, Kansas; Mudge (1955) believed that it would run northeast-southwest from a position one mile northwest of St. George until it would join with Rock Creek valley west of Louisville.

The main feature of the topography of the bedrock surface is the asymetric, buried, pre-glacial valley (Fig. 10). The buried valley trends east-northeast from an area approximately one half mile west of St. George to an area slightly over four miles north of St. George. The buried valley trends east-west for approximately four and a half miles until it merges with the valley of Rock Creek slightly south of Louisville. Rock Creek valley trends to the southeast for approximately three miles until it joins the Vermillion Creek valley.

The gradient of the buried, pre-glacial valley is approximately 5 feet per mile as it trends to the north-northeast and to the east (Fig. 10). The buried, pre-glacial valley varies between one and two miles in width as it trends to the north-northeast, and its width increases to between two and three miles as it trends to the east. The depth of the buried, pre-glacial valley below the valley walls varies between 25 and 175 feet in depth. The elevation of the buried, pre-

38
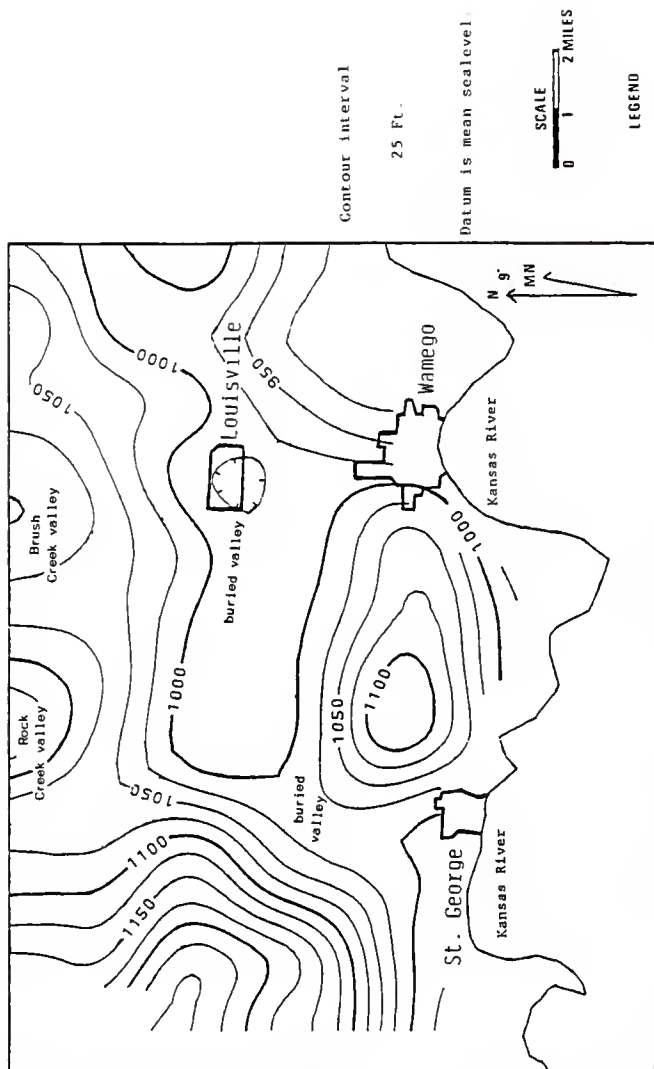
Figure 10. Elevation of the bedrock in southwest Pottawatomie County, Kansas.

Contour interval

25 Ft.

Datum is mean sealevel

SCALE

0   1   2 MILES

LEGEND

N 9° MN

Louisville

Wamego

Kansas River

Brush Creek valley

buried valley

Rock Creek valley

buried valley

St. George

Kansas River

1000

1050

950

1000

1100

1050

1000

1050

1100

1150

glacial valley is between 975 and 1000 feet for most of its length,
and it is very similar to the elevation of the modern Kansas River
floodplain.

Stratigraphic correlation diagrams of the buried, pre-glacial val-
ley were constructed from information obtained from water well logs
(Fig. 11; Appendix VII). The upper deposits of the buried, pre-glacial
valley are chiefly clay and sandy clay (Fig. 12; 13). The deposits
grade downward to fine sand, and toward the center of the buried, pre-
glacial valley, coarse sand and gravel. In general, the number of fin-
ing sequences in the buried, pre-glacial valley becomes greater as the
sides of the valley are approached, and the grain size of the sedi-
ments increases as the center of the buried, pre-glacial valley is ap-
proached. The individual fining upward sequences of the buried, pre-
glacial valley range in thickness from less than ten feet to a thick-
ness of over 120 feet. The individual fining upward sequences of the
Kansas River valley commonly have thicknesses of 50 feet or less.

Chert gravels of variable thickness occur in the buried valley to
the south and west of Rock Creek. The gravels are valuable aquifers in
the study area. The thickness of the gravels have been contoured
(Fig. 14). The thickest deposit of gravels is 30 ft thick, and it is
located 1.5 miles to the north-northwest of Wamego.

Hydrogeology of the Hydrologic Units

The geologic units of the area have been divided into the buried
valley sediments, Kansas River alluvium, upland stream alluvium, Qua-
ternary sediments, and bedrock hydrologic units to facilitate data
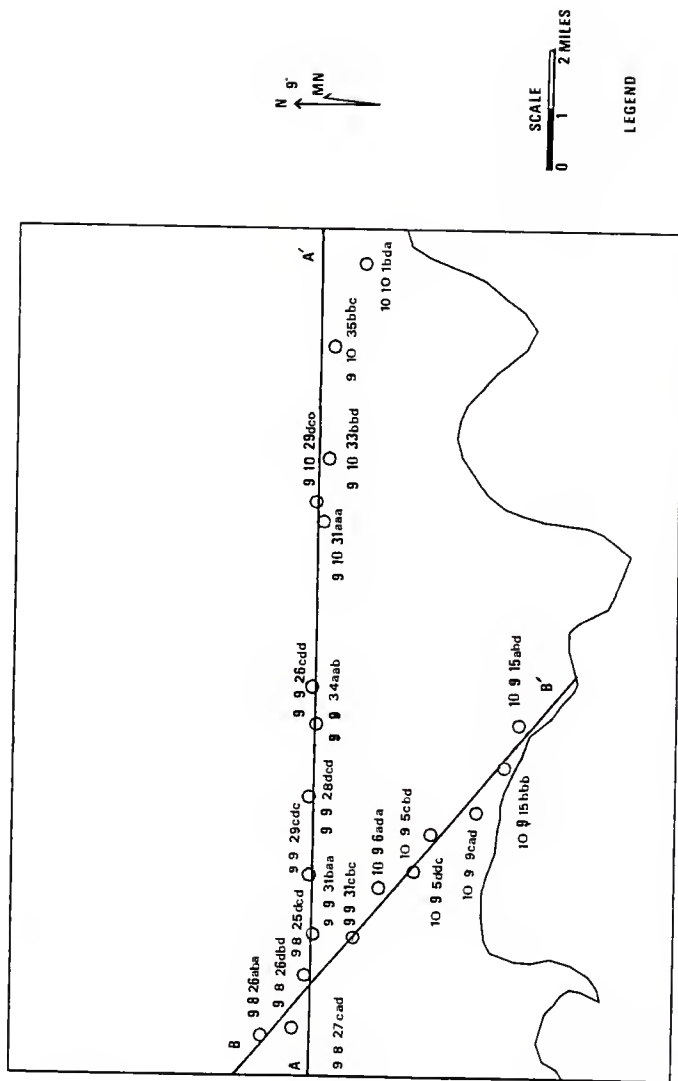
Figure 11. Location of stratigraphic correlation diagram lines and the wells used to construct the stratigraphic correlation diagrams.
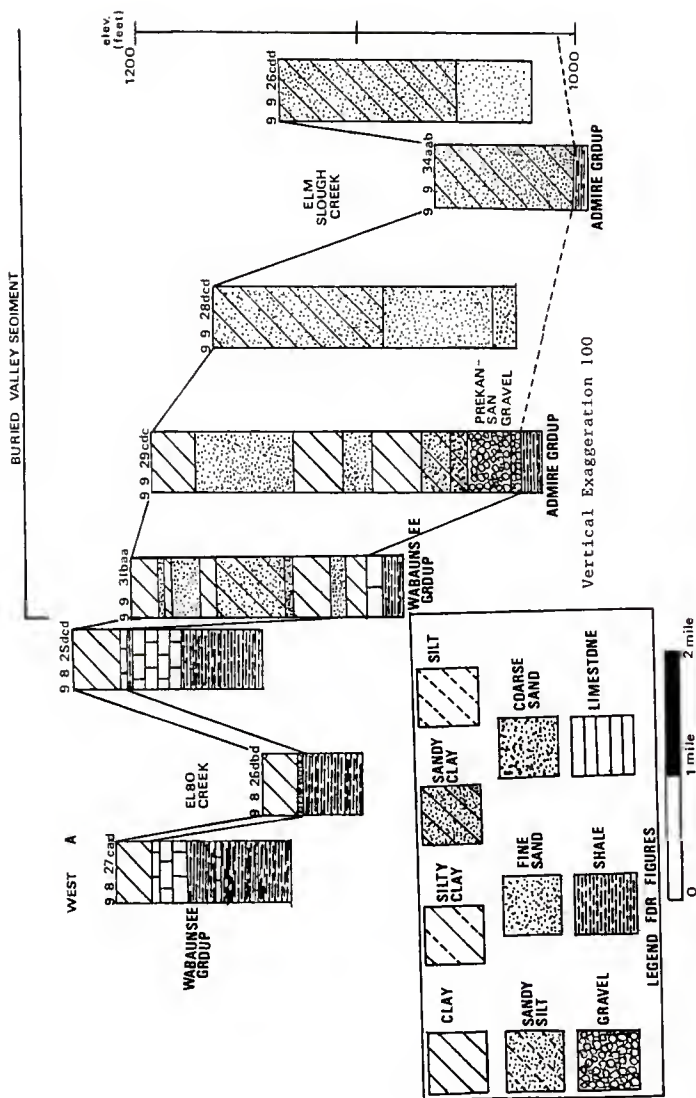
Figure 12. Stratigraphic correlation diagram A-A. (See Figure 11 for the location of the wells.)

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Figure 12. continued

Vertical Exaggeration 100

43
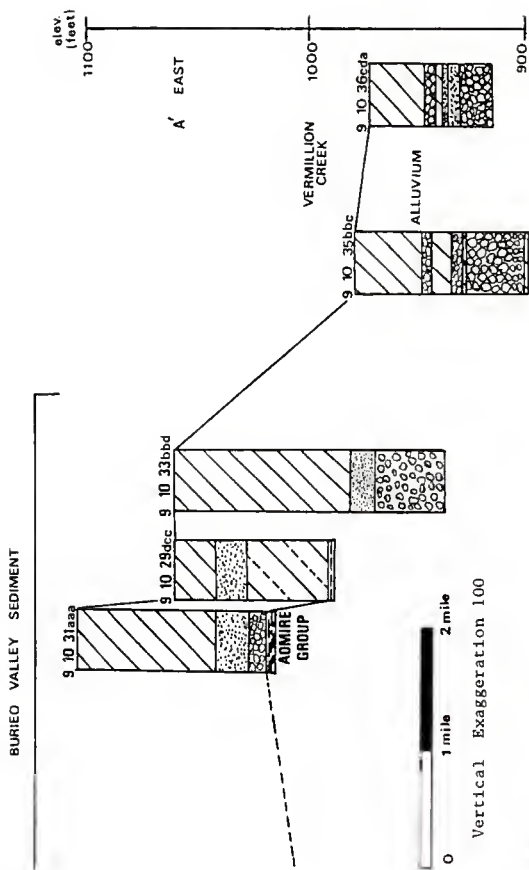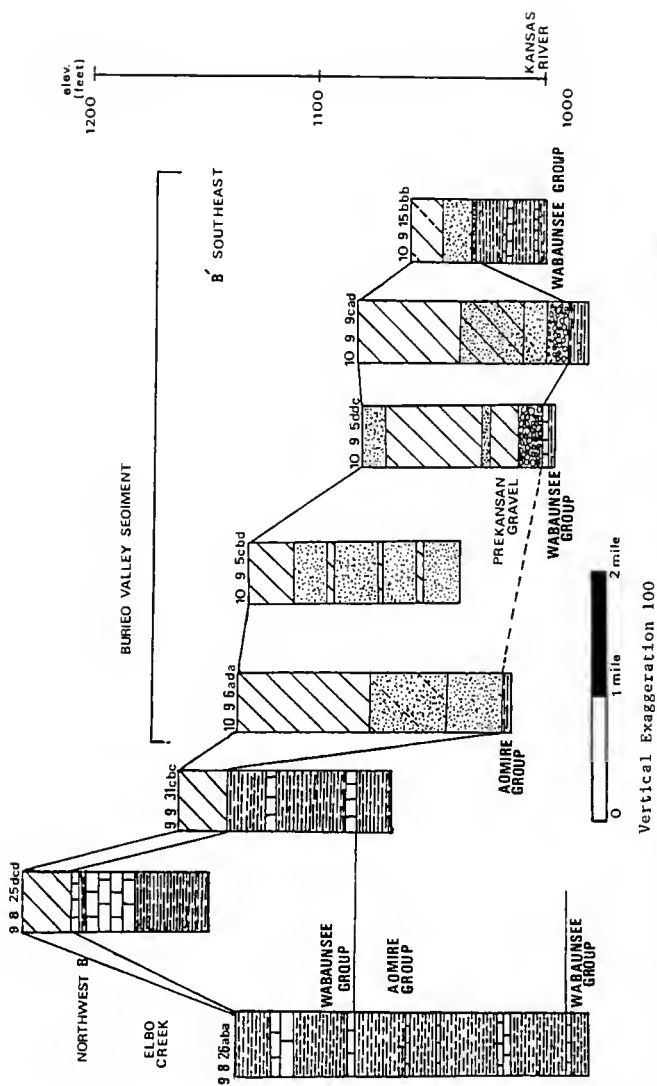
Figure 13. Stratigraphic correlation diagram B-B'. (See Figure 11 for the location of the wells and Figure 12 for legend.)

44
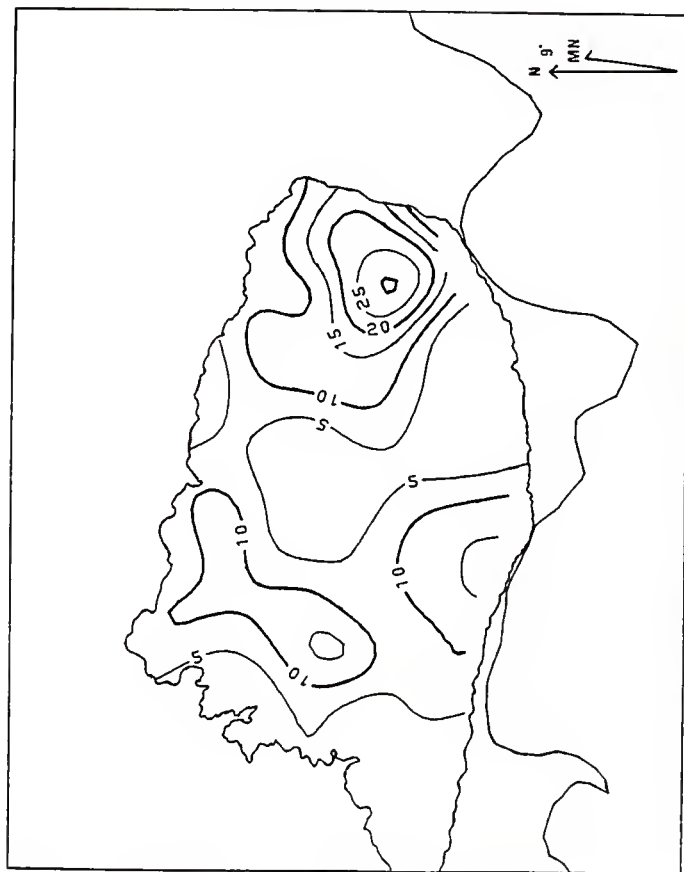
contour interval
5 Ft.

SCALE

0   1   2 MILES

LEGEND

Figure 14. The thickness of the chert gravel deposits found within the buried valley.

analysis. The hydrologic units are similar to the divisions used by Walters (1953; 1954), the Kansas Water Resources Board (1959), Whittemore and others (1982) and Stephen Farley (personal communication, 1983).

Buried Valley Sediments--The buried valley sediments consist of the glacio-lacustrine, glacio-fluvial and pre-Kansan gravel deposits that have filled the buried, pre-glacial valley to the south and to the west of Rock Creek.

The permeability and groundwater yield of the buried valley sediments differs with grain size (Table 4; O, Conner, 1971; Walters, 1953; 1954). Water wells that penetrate the permeable sands and gravels in the bottom of the buried valley have large yields of water and may produce up to 450 gallons of water per minute (Table 4; Appendix VIII). The silty sands and clayey silts produce less than 40 gallons of water per minute because of the small grain size of the deposits (Appendix IX; O'Conner, 1971).

The depth to water in the buried valley ranges geographically from 17 ft (6 m) to 98 ft (32 m) below the surface (Table 4). In general, the gravel deposits are water saturated, but only a portion of the sand and silt deposits are saturated. An impermeable layer may overlie a sloping water-bearing layer and confine the water in the permeable layer (Walters, 1953; 1954). Artesian pressure is often observed in wells that obtain water from the buried valley sediments. The dominant clay mineral of the buried valley sediments is illite (Snow, 1963).

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Table 4. The average, low, and high values for the depth to the watertable and the yield of wells in the hydrologic units in the area of investigation (Appendix VII).

Hydrologic Unit	Characteristic	Mean	Low	High	Number of Samples
Bedrock	Depth (Ft.)	49.9	19	172	20
	Yield (GPM)	11.4	.5	30	19
Glacial Deposits	Depth (Ft.)	33.1	12	80	7
	Yield (GPM)	23.6	1.5	80	4
Kansas River Alluvium	Depth (Ft.)	26.9	11	54	31
	Yield (GPM)	571.9	6	3000	30
Upland Stream Alluvium	Depth (Ft.)	29	12	40	9
	Yield (GPM)	158.4	4	1200	8
Buried Valley Sediments	Depth (Ft.)	55.5	17	98	64
	Yield (GPM)	43.3	3	450	67

<u>Upland Stream Alluvium</u>--The upland stream alluvium includes the flood-plain alluvium and terrace deposits of all the major streams of the area except the Kansas River.

The alluvium of upland streams is a major source of groundwater in northeast Kansas (Frye and Walters, 1950). The soils of the upland stream alluvium have moderate to low permeability and the dominant clay mineral of the upland stream sediments is montmorillite (Smec-tite; Soil Conservation Service, 1975; 1980; Switek, 1977). The sands and gravels of the upland stream alluvium are highly permeable and the yields of the upland stream alluvium range from 4 to 1200 gallons per minute (Table 4). In general, the wider the valley and the thicker the sediments, the larger the groundwater yields of the alluvium (Walters, 1953; 1954).

The upland stream alluvium is thin and ranges in thickness from a feather edge to 40 ft (13 m). The depth to water in wells that are not being pumped is usually shallow and averages 29 feet (Table 4). Water in the upland stream alluvium is usually unconfined (Scott and others, 1959; Frye and Walters, 1950).

<u>Kansas River Alluvium</u>--The Kansas River alluvium consists of the terrace deposits and floodplain alluvium of the Kansas River.

The Kansas River alluvium is the most productive aquifer in the area. The Kansas River floodplain has soils with moderate to low per-meability, and the dominant clay mineral of the Kansas River Alluvium is montmorillite (smectite; Soil Conservation Service, 1975; 1980; Switek, 1977). The sands and gravels of the Kansas River alluvium are

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very permeable and have yields that may be as high as 3000 gallons per minute and average 572 gallons per minute (Table 4).

The depth to groundwater in the Kansas River alluvium ranges geographically from 11-54 ft (3-18 m) and the thickness of the valley fill may be as great as 100 ft (33 m; Table 4; Fader, 1974; Beck, 1959; Smith, 1959; Davis and Carlson, 1952). The high yields and the shallow water table of the Kansas River alluvium allows the Kansas River valley to be extensively irrigated (Scott and others, 1959).

Glacial Deposits--The glacial deposits are made up of the glacial deposits that are not in the buried valley and they include fluvial deposits, lacustrine deposits, glacial till, and loess.

The amount of water from the glacial deposits varies greatly because of the great range in the lithology of the glacial deposits. Loess and glacial till are fine-grained sediments with low permeability, and wells obtaining water from them have low yields (Table 4; Walters, 1953; 1954). Sand and gravel deposits that are interbedded with the glacial till have high permeability, but often fail as aquifers because of a lack of sufficient area for recharge (Frye and Walters, 1950; Heath, 1984). The average yield of 4 gallons per minute reflects the small average grain size of the glacial deposits (Table 4).

The soils that form on the glacial till and loess have moderately low to very low permeability. Soils forming on glacial outwash have moderately high permeability (Soil Conservation Service, 1975; 1980). The dominant clay of the glacial aquifers is montmorillite (smectite), but some glacial outwash aquifers have illite as their dominant clay

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mineral (Freeze and Cherry, 1978; Snow, 1963).

The permeability and recharge of the glacial deposits may be increased upto a hundred times by jointing (Williams and Farvolden, 1967). Joints occur in some of the glacial deposits of Pottawatomie County (Frye and Walters, 1950; Scott and others, 1959). Jointing of the glacial deposits explains why some glacial deposits have relatively rapid changes in water quality due to changes in precipitation (Hendry, 1982; Sharp, 1984).

The depth to water in wells in the glacial deposits is from 12 to 80 ft (4-13 m; Table 4). Glacial deposits range in thickness from a feather edge to over 80 ft (13 m; Scott and others, 1959; Beck, 1959). Many wells in glacial sediments have artesian pressure and in some areas a perched water table exists where water-bearing glacial deposits are underlain by impermeable and non-water bearing permeable sediments (Smith, 1959; Walters, 1953; 1954).

Bedrock--The bedrock hydrologic group is composed of limestone, shale, and sandstone of Paleozoic age.

The bedrock aquifers have low permeability and small to moderate yields (Walters, 1954; 1953; Beck, 1959). The soils that form over shale are deep and have moderately low to very low permeability. The soils that form over limestone are shallow soils with moderately low permeability (Soil Conservation Service, 1975; 1980; Switek, 1977). The most abundant clay mineral in the bedrock is illite (Twiss, 1955; Dulekoz, 1966; Asmussen, 1958; Dowling, 1967; Hargedine, 1963; Voss, 1972).

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The limestones have had their porosity and their yields increased by jointing. The solution of the limestone along the joints and bedding planes has resulted in the formation of channels and cavities for the movement and storage of water. The effects of solution and the yield are greatest near the outcrop of the limestone (Whittemore and others, 1982; Heath, 1984; Langmuir, 1971). Wells in the bedrock have an average yield of 11 gallons per minute (Table 4).

The depth to water in wells that are not being pumped changes greatly geographically. The great differences in the elevation of the different bedrock units has caused the depth to water in wells in the bedrock to range from 19 to 172 ft (6-44 m; Table 4). Water obtained from the bedrock is often under artesian pressure, particularly in the areas between the uplands and major stream valleys, but there are no flowing wells in the study area (Smith, 1959; Heath, 1984).

Recharge and Discharge of the Hydrologic Groups--Recharge is the addition of water to the zone of saturation within the aquifer whereas, discharge is the removal of water from the zone of saturation within the aquifer. Some means of recharge are also means of discharge, and some processes affect both recharge and discharge (Todd, 1980).

Sources of recharge for the aquifers are flow of water from another aquifer, infiltration from streams, ponds, and lakes to aquifers, and infiltration of rainwater (Beck, 1959; Smith, 1959). Infiltration of rainwater is the major source of water for most of the aquifers (Walters, 1953; 1954; Beck, 1959; Smith, 1959). The period for the greatest recharge is biseasonal, occurring mostly in late Fall and

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early Spring when the ground is thawed and rainfall is greatest (O'Conners, 1972; Heath, 1984). The amount of time that it takes for water to infiltrate to the zone of saturation varies (Whittemore and others, 1982). The dilution effect of periods of great rainfall may be seen in bedrock aquifers within six months, but may not be seen in glacial aquifers that have long periods of recharge (Whittemore and others, 1982).

The aquifers of the area receive very little recharge from ponds, lakes or streams. The streams are influent streams that receive water from the alluvium (Davis and Carlson, 1952; Beck, 1959). The ponds and lakes that are built in the uplands have clay lined bottoms that inhibit the infiltration of water into the underlying aquifers (Smith, 1959). An unusual form of recharge called bank full storage may occur when the water level of a stream is higher than the level of the zone of saturation in the alluvium and stream water flows into the alluvium (Beck, 1959; Smith, 1959). The level of water in wells near the Kansas River mirror the water level of the Kansas River with only a small time lag and cause the water table in the Kansas River alluvium to vary by as much as 10 to 15 ft (3-5 m; Davis and Carlson, 1952; Beck, 1959; Smith, 1959).

Discharge occurs by wells, flow into another aquifer, springs and seeps, regional flow, evaporation, and transpiration (Walters, 1953; 1954; Smith, 1959; Beck, 1959).

Evaporation and transpiration are processes that affect both infiltrating water and water in the zone of saturation. Evaporation causes discharge only if the water table is within a few feet of the

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land surface (Walters, 1953; 1954). Evaporation is temperature dependant, and it will increase as the surface temperature increases (O'Conners, 1972). Evaporation occurs in the soil zone, streams, ponds, lakes, marshes, and, occasionally, alluvium in the area of investigation (Smith, 1959). Transpiration is the process by which plants absorb water from the soil, use it, and give it off to the atmosphere. Transpiration is most common during the summer when vegetation is abundant and active (O'Conner, 1971). Discharge by transpiration gets greater as the zone of saturation gets closer to the surface, and is greatest when the roots of grasses, which may be up to 20 ft long can reach the zone of saturation (Twiss, 1987, personnal communication; Walters, 1953; 1954). The water tables of the Kansas River alluvium, upland stream alluvium and Quaternary sediments may be discharged by transpiration (Walters, 1953; 1954; Beck, 1959).

Water reaching the zone of saturation moves down gradient until it reaches a point where it leaves the aquifer (Heath, 1984). Aquifers that are down gradient from another aquifer will often receive recharge from the aquifer above it (Walters, 1953; 1954). Water flows from upland deposits into alluvium, and it will eventually arrive at the Kansas River alluvium (Whittemore and others, 1982). Water in the Kansas River alluvium will move down stream (Beck, 1959).

Small amounts of water, usually from bedrock aquifers, will flow into the glacial aquifers from other aquifers because of the small grain size of the glacial aquifers (O'Connor, 1971; Heath, 1984). Water flowing into glacial aquifers from bedrock aquifers contains high total dissolved solids because the slow movement of water into

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the glacial aquifer allows the water to dissolve large amounts of
materials from the bedrock (Walters, 1953; 1954; Heath, 1984).

The flow of water into bedrock is usually small because the small
grain size of the shales in the bedrock inhibits the infiltration of
water into the bedrock (Todd, 1980; Heath, 1984). The shales inhibit
the movement of water from one bedrock aquifer to another. In general,
any water that flows into the bedrock will flow into limestone and not
shale (Walters, 1953; 1954). The Paleozoic bedrock of Pottawatomie
County is highly dissected by streams, and a bedrock aquifer is un-
likely to be able to carry water far in such terrain (Davis and Carl-
son, 1952). Water in the Paleozoic bedrock will flow down dip out of
the area or into the valley of a major stream (Chelikowsky, 1972;
Gregory, 1963).

The Kansas River and upland stream alluvium are permeable and re-
ceive large amounts of recharge from other aquifers (Smith, 1959;
Whittemore and others, 1982). (Smith, 1959; Beck, 1959). The upland
aquifers either come directly into contact with alluvium or they are
underlain by an impermeable layer and discharge their water as a
spring (Walters, 1953; 1954; Whittemore and others, 1982). The amount
of recharge from the other aquifers is large enough to enable some
streams to flow all year round (Davis and Carlson, 1952; Whittemore
and others, 1982). Bedrock springs occur when a permeable limestone
bed is underlain by a shale and they are especially common in bedrock
exposures in the north walls of the Kansas River and Rock Creek val-
leys (Beck, 1959; Smith, 1959). In the low areas of the buried valley
and the glacial deposits, the ditches and the drainage ways often have

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flowing water from seeps. The seeps of the buried valley and the gla-
cial deposits occur where the downward movement of infiltrating water
is blocked by an impermeable layer, and the water is discharged on to
the land surface (Walters, 1953; 1954).

Chemical Results

The results of the chemical analyses of the water samples are
given in Appendix III and summarized in table 5. Chemical data of
previous analyses of groundwater from the area of investigation are
shown in Appendix I.

Calcium concentrations in the groundwater of the area are high
because large amounts of soluble calcium-bearing minerals occur in the
area (Table 5). The calcium of the sample waters may be derived from
the dissolution of calcite, dolomite, gypsum, and minor amounts of an-
hydrite (Beck, 1959; Smith, 1959).

Magnesium concentrations in the groundwater of the area are lower
than the calcium concentrations (Table 5). The most common source of
magnesium in sedimentary terranes is the dissolution of dolomite and
calcite, both of which occur in the area (Twiss, 1955; Dulekoz, 1966;
Watkins, 1957; Schmidt, 1974; Dowling, 1967; Asmussen, 1958; Harga-
dine, 1963). The dissolution of pyroxenes and amphiboles may add small
amounts of magnesium to water in the glacial deposits (Hem, 1970).

The sodium concentrations of the groundwater of the area approxi-
mately equal the magnesium concentrations (Table 5). Marine water
trapped in pore space in sediments, ion exchange with clay minerals,
and the dissolution of halite are the common natural sources of sodium

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Table 5. The mean, low, and high concentrations of the chemical constituents of the water in southwest Pottawatomie County, Kansas.

	Mean	Low	High	Number of Samples
Calcium	94	10	324	70
Magnesium	14	.7	39	70
Sodium	30	1.6	140	70
Potassium	3.7	.7	28	70
Strontium	1.6	DL	26	70
Iron	.8	DL	13.8	70
Manganese	.16	DL	2.4	70
Bicarbonate	280	57	511	70
Sulfate	90	DL	900	70
Chloride	16	1.0	103	70
Nitrate	31	DL	220	40

Note: All concentrations are expressed as milligrams per liter.

Note: All specific electrical conductivities are expressed as micromhos per centimeter squared.

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in water from sedimentary terranes (Hem, 1970; Matthess, 1982). The weathering of feldspar may provide a small amount of the sodium in the water of the buried valley sediments (Snow, 1963; Hem, 1970; Erikson, 1985). Common man-made sources of sodium are sewage, industrial wastes, animal wastes, and road salt (Hem, 1970; Lloyd and Heathcote, 1985).

The potassium concentrations in the ground water of the area have a large range (Table 5). The most common source of potassium in sedimentary terranes are fertilizers (Matthess, 1982; Erikson, 1985). Natural sources of potassium are potassic micas and clay minerals (Hem, 1970; Matthess, 1982).

The iron, manganese, and strontium concentrations in the groundwater of the area of investigation have large ranges (Table 5). The principal natural sources of iron in water are hematite and limonite (Hem, 1970). Iron carbonates and sulfides may also contribute small amounts of iron to water. The corrosion of metal parts of water wells give the well waters high iron concentrations (Hem, 1970; Lloyd and Heathcote, 1985). The principal source of manganese in water is organic matter in soils. Manganese will also substitute for iron in many iron bearing minerals (Hem, 1970; Erikson, 1985). The natural sources of strontium in water are celestite, strontianite, calcite, and aragonite (Hem, 1970). Strontium may occur as an impurity in fertilizers.

The molal concentration of bicarbonate is greater than the molal concentration of any other ion in nearly all the sample waters (Table 5). The sources of the bicarbonate in water are the solution of carbon dioxide from the air, the solution of carbon dioxide from the soil,

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and the solution of calcite and dolomite (Hem, 1970).

The concentration of sulfate varies a great deal in the area of investigation (Table 5). The dissolution of gypsum and anhydrite are the principal sources of sulfate in water (Hem, 1970; Brickers and Garrels, 1968). Celestite dissolution and the oxidation of sulfides, chiefly pyrite, may add small amounts of sulfate to water (Whittemore and Switek, 1977; Hem, 1970; Lloyd and Heathcote, 1985).

The chloride and nitrate concentrations in the groundwater of the area are small and have large ranges (Table 5). The natural sources of chloride in water are marine waters trapped in the pore spaces of se-diments and the dissolution of chloride salts (Hem, 1970; Davis and Dewiest, 1966). Chloride may also come from highway salts, fertili-zers, sewage, and animal wastes (Hem, 1970; Lloyd and Heathcote, 1985). Natural sources of nitrate are the decay of large amounts of plant material, the growing of legumes in an area of recharge, and animal wastes. Man made sources of nitrate are fertilizers and sewage (Hem, 1970; Matthess, 1982).

The average water sample of the area is calcium-bicarbonate type water (Table 5; Hem, 1970). Eight of the seventy samples did not have calcium-bicarbonate type water. Those wells that were not calcium-baicarbonate type water were primarily mixed types of water. Thirty of the seventy water samples taken had total dissolved solids above the United States Public Health Service standard (U.S.P.H.S.) of 500 mg/l.

DISCUSSION

All the chemical results were used to characterize the chemical composition, the saturation indices, and the seasonal variation of the groundwater. The first sample series were the only analyses used to determine the geographical variation of the groundwater of the area.

Saturation Indices

The saturation index (S.I.) of a mineral indicates the degree of saturation of a solution with respect to the mineral. The saturation index of a mineral is calculated according to the equation (Hem, 1985),

$$SIa = \log \frac{IAPa}{Ka} \qquad (6)$$

where, SIa = saturation index of the solution with respect to solid A.

    IAPa = Ion activity product of the ions of solid A in solution

    Ka = Solubility constant of solid A.

A saturation index equal to zero indicates that a solution is saturated with respect to the solid (i.e. in equilibrium with that solid). A saturation index less than zero indicates a solution that is undersaturated with respect to the solid and that is capable of dissolving more solid. A saturation index greater than zero indicates a solution that is supersaturated with respect to a solid and capable of precipitating the solid (Hem, 1970; Krauskopf, 1979).

The solubility constant of a solid is determined experimentally, and the ion activity product of a solid in solution is calculated using an equation that is based on the equation that describes the dissolution of the solid. The general form of such equations and the ion activity product of the reaction are as follows '(Krauskopf, 1979; Hem, 1970):

$$AdBc = dA + cB \qquad (7)$$

$$IAP = \gamma_{a(mA)}^{d} \; \gamma_{b(mB)}^{c} \qquad (8)$$

where, IAP = the ion activity product of the solid in solution.

    mA = the concentration of ion A in

        solution in moles per liter.

    mB = the concentration of ion B

        in solution in moles per liter.

    γ_a = the activity of ion A in solution.

    γ_b = the activity of ion B in solution.

    d = the coefficient of ion A.

    c = the coefficient of ion B.

The effect of other ions in solution on a particular ion is described by the activity of the ion in the solution. The activities of ions in the sample waters were determined from a graph showing ionic activities in water as a function of the ionic strength of the water (Hem, 1985, Fig. 1). The activity coefficients of the graph were cal-

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culated using the Debye-Heckel equation (Hem, 1985),

$$-\log \gamma = \frac{A z^2 \sqrt{I}}{1 + Ba \sqrt{I}}$$ (9)

where, γ = the activity of the ion in the solution.

   A = a constant to account for the viscosity of

        the solvent, water = .5085.

   B = a constant for water, .3281.

   z = the charge of the ion.

   a = a constant for the diameter of the ion.

   I = the ionic strength of the solution.

The ionic strengths of the sample waters (Table 7; Appendix IX) were calculated using the equation,

$$I = 0.5 \ ((C)(z^2))$$ (10)

where, I = ionic strength of the water.

   C = molarity of ionic species

        in solution.

   Z = ionic charge of ionic species

        in solution.

Only the S.I. of calcite, dolomite, and gypsum were calculated

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for the sample waters because the dissolution of these minerals appear to provide most of the dissolved solids of the sample waters, and they control the upper limits of the calcium, magnesium, bicarbonate, and sulfate concentrations of the sample waters (Hem, 1970; Bricker and Garrels, 1968). Equations 11, 12, and 13 describe the equilibrium of dissolved calcite, dolomite, and gypsum in waters (Table 6).

Carbonate was calculated using the known bicarbonate and hydrogen ion concentrations and the following relation:

$$HCO_3^- = H^+ + CO_3^{2-} \qquad K_h = 10^{-10.4} \qquad (14)$$

$$K_h = 10^{-10.4} = \frac{(mH) \, \gamma co_3 (mCC_3)}{\gamma \, co_3 \, (mHCO_3)} \qquad (15)$$

or

$$\gamma co_3 (mCO_3) = \frac{10^{-10.4} \gamma \, co_3 (mHCO_3)}{mH} \qquad (16)$$

where, K_h = the equilibrium constant for the
  dissociation of bicarbonate.

$\gamma \, co_3$ = the activity coefficient of bicarbonate in solution.

  γH = the activity coefficient of hydrogen in solution.

γco_3 = the activity coefficient of carbonate in solution.

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Table 6. The equations, solubility constants, and ion activity products of calcite, dolomite, and gypsum used in the study (Krauskopf, 1979).

Mineral	Equation	Solubility constant	Ion activity product (IAP)	
Calcite	$CaCO_3 = Ca^{2+} + CO_3^{2-}$	$10^{-8.22}$	$\gamma_{ca}(mCa^{2+})\ \gamma_{co_3}(mCO_3^{2-})$	(11)
Dolomite	$CaMg(CO_3)_2 = Ca^{2+} + Mg^{2+} + 2CO_3^{2-}$	$10^{-17.2}$	$\gamma_{ca}(mCa^{2+})\ \gamma_{mg}(mMg^{2+})\ \gamma_{co_3}(mCO_3^{2-})^2$	(12)
Gypsum	$CaSO_4 \cdot nH_2O = Ca^{2+} + SO_4^{2-} + nH_2O$	$10^{-4.6}$	$\gamma_{ca}(mCa^{2+})\ \gamma_{so_4}(mSO_4^{2-})$	(13)

where, γ_{ca} = activity coefficient of calcium in solution

γ_{mg} = activity coefficient of magnesium in solution

γ_{so_4} = activity coefficient of sulfate in solution

γ_{co_3} = activity coefficient of bicarbonate in solution

mCa^{2+} = molarity of calcium in solution

mMg^{2+} = molarity of magnesium in solution

$mHCO_3^-$ = molarity of bicarbonate in solution

mSO_4^{2-} = molarity of sulfate in solution

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$mHCO_3$ = the molarity of bicarbonate in solution.

mCO_3 = the molarity of carbonate in solution.

mH = the molarity of hydrogen in solution.

Thirteen out of seventy water samples were saturated or supersaturated with respect to calcite (Appendix IX). The water samples from the Kansas River alluvium had the highest average calcite S.I. (Table 7). The water samples from the buried valley sediments had the lowest average calcite S.I., -2.94, and a median calcite S.I. value of -1.07. Only one sample from the buried valley deposits was saturated with respect to calcite. Water samples from the buried valley in Marshall County had calcite S.I. with a median value of -.05, and appear to have contained approximately ten times more dissolved calcite than water samples from the buried valley sediments in southwest Pottawatomie County (Farley, personal communication, 1987).

Fifteen of the water samples were saturated or supersaturated with respect to dolomite (Table 7). The water samples from the Kansas River alluvium had the highest average dolomite saturation index, -.053. Water samples from the buried valley had the lowest dolomite S.I. and average dolomite S.I., -1.90, and a median dolomite S.I. value of -2.19. The buried valley in Marshall County had water samples with a median dolomite S.I. value of -.67, and appeared to have approximately 30 times more dissolved dolomite than the water samples from the buried valley sediments in Pottawatomie County.

The water samples from the Kansas River alluvium, bedrock, and all but one of the water samples from the glacial deposits have cal-

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Table 7. The mean, low, and high values of the ionic strength and the saturation indices of the waters from the hydrologic units in Pottawatomie County and the buried channel in Marshall County (Farley, 1988).

	Saturation Indices of Calcite			Saturation Indices of Dolomite			Saturation Indices of Gypsum			Ionic Strength (Moles per Liter)			Number of Samples
	Mean	Low	High	Mean	Low	High	Mean	Low	High	Mean	Low	High	
Bedrock	-.193	-.435	-.008	-.088	-.858	.277	-1.64	-2.66	-.74	.01359	.00566	.02556	10
Buried Valley	-0.83	-2.39	-.076	-1.90	-4.95	.172	-2.77	-3.33	-1.36	.00612	.00183	.00813	30
Glacial Deposits	-.178	-.755	.476	-.263	-1.63	.945	-.77	-1.43	.24	.02271	.04217	.01378	10
Kansas River A.	-.008	-.285	.314	-.053	-.488	.515	-1.62	-2.44	-.90	.01270	.00809	.01859	10
Upland Stream A.	-.305	-1.36	.413	-.550	-2.73	.959	-1.99	-3.19	-.57	.00979	.00228	.01387	10
All Samples	-1.21	-2.39	.476	-.943	-4.95	.959	-1.95	-3.19	.24	.01131	.00183	.02556	10
Marshall County	-0.05	-.15	.10	-.67	-1.05	-.51	-2.03	-2.51	-1.58				10

Note: The mean values listed for the buried channel in Marshall County are actually median values.

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cite and dolomite saturation indices greater than -1.0 (Fig. 15). The calcite and dolomite saturation indices of the upland stream alluvium vary greatly. The calcite and dolomite saturation indices with the largest range are those of the buried valley sediments. The average calcite saturation index of the buried valley sediments, -2.94, is an order of magnitude lower than the buried valley sediments average dolomite S.I. -1.90.

Most samples form a linear array in which the dolomite saturation index increases at a slightly greater rate than the calcite saturation index (Fig. 15). The precipitation of calcite is probably not the cause of the different rates of increase because the rate of increase of the saturation indices is constant. The different rates of change of the S.I. is probably due to the dissolution of dolomite (Hargadine, 1963; Asmussen, 1958; Dowling, 1967; Dulekoz, 1966; Twiss, 1955; Schmidt, 1974; Watkins, 1957).

Some samples are supersaturated with respect to both calcite and dolomite (Fig. 15). Magnesium increases the free energy of the calcite molecule, and the presence of large amounts of magnesium in solution may cause the calcite solubility to become greater than the calcite solubility predicted by the calcite S.I. (Bricker and Garrels, 1968; Matthess, 1985). Gypsum dissolution, or recharge by waters containing large amounts of dissolved gypsum, may also cause water to become supersaturated with respect to calcite, because the dissolution of gypsum appears to occur more quickly than the precipitation of calcite (Hem, 1970; Brickers and Garrels, 1968).

The maximum dolomite saturation index of the sample waters is

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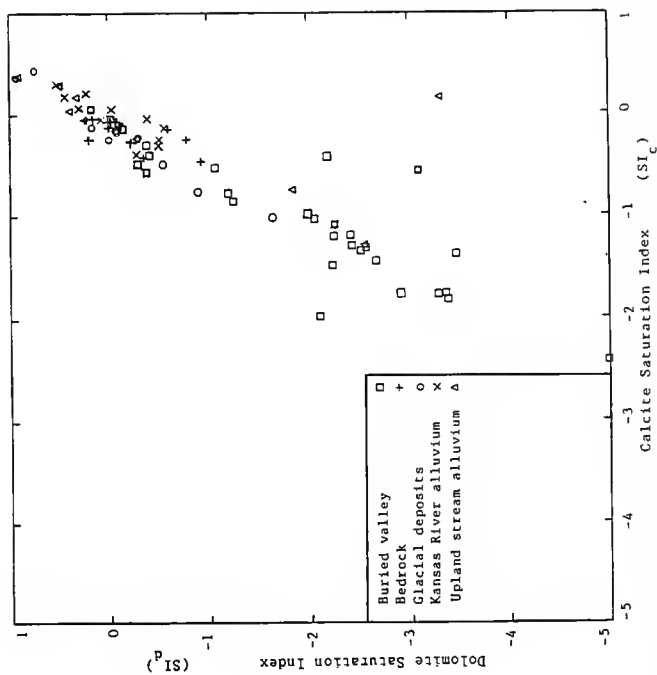
Figure 15. The calcite saturation index-dolomite saturation index relationships of groundwater in southwest Pottawatomie County, Kansas.

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much higher than the maximum calcite saturation index (Fig. 15). The supersaturation of water with respect to dolomite may be caused by a tendancy for calcite to precipitate out of solution more easily than dolomite (Hem, 1970; Bricker and Garrels, 1968; Langmuir, 1971). Water that is in contact with dolomite may become supersaturated with respect to dolomite as calcite precipitates and dolomite dissolves (Bricker and Garrels, 1968; Langmuir, 1971). The mixing of water containing large amounts of dissolved gypsum and water supersaturated with respect to calcite or dolomite will cause calcite to precipitate and supersaturation with respect to dolomite (Whittemore, personal communication, 1988).

Only two water samples, both from the glacial deposits, are saturated or supersaturated with respect to gypsum (Appendix IX). The average gypsum S.I. is lower than the average calcite and dolomite S.I. (Table 7). The glacial deposits had the water samples with the highest average gypsum saturation index, -.77. Water samples from the buried valley sediments had the lowest average gypsum S.I. and a median gypsum S.I. of -2.83. Water samples from the buried valley in Marshall County had gypsum S.I. with a median value of -2.03, and appear to have contained approximately six times more dissolved gypsum than the water samples from the buried valley in Pottawatomie County.

When the S.I. of calcite and gypsum of the sample waters are plotted against each other most of the points follow two trends (Fig. 16). One trend is formed by samples with low gypsum S.I. and highly variable calcite S.I. Many of these samples are from the buried valley sediments where nearly all the gypsum S.I. are less than -2.0. The

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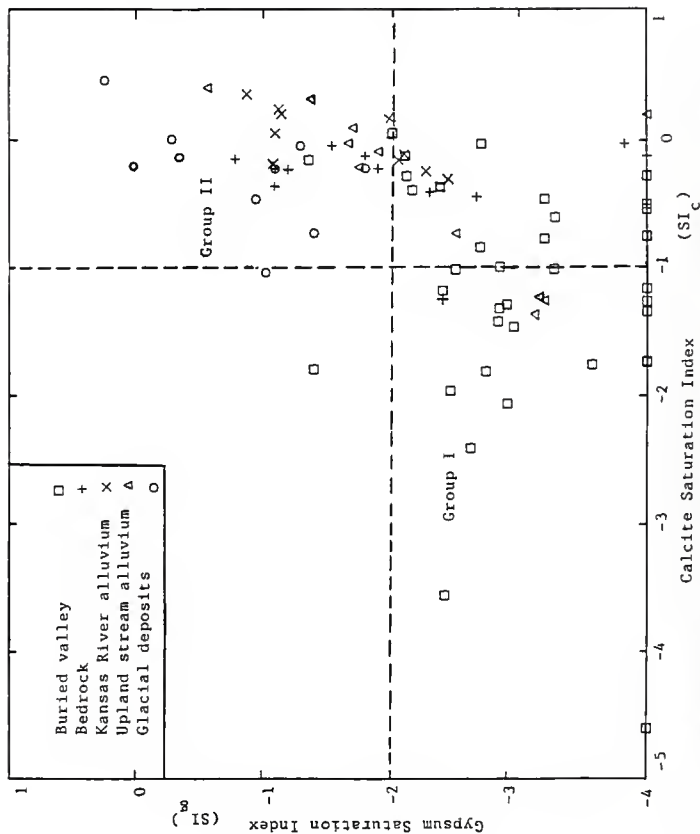
Figure 16. The calcite saturation index-gypsum saturation index relationships of groundwater in southwest Pottawatomie County, Kansas.

second trend is formed by samples with high calcite S.I. that often range between .5 and -.5, and highly variable gypsum S.I. that range between -2.0 and .25 (Fig. 16). In general, the samples with the highest gypsum S.I. tend to have the highest calcite S.I. The dissolution of gypsum and the precipitation of calcite can explain the trend of the second group of samples. Gypsum dissolution or recharge by water containing dissolved gypsum may increase the calcium concentration of water already high in calcium and cause the waters to become supersaturated with respect to calcite. Over time, calcite will precipitate from the solution, and gypsum will continue to dissolve and increase in concentration (Langmuir, 1971; Lloyd and Heathcote, 1985).

Two water samples from the glacial deposits are supersaturated with respect to gypsum and calcite (Fig. 16). The supersaturation of waters with respect to gypsum and calcite may be caused by the mixing of water from the bedrock containing dissolved gypsum with water in the glacial deposits containing dissolved calcite (Hem, 1970). The mixing of waters could also explain the high concentrations of gypsum in the upland stream alluvium and the Kansas River alluvium.

Ion Ratios

Magnesium-Calcium Ratio--The line labeled "meq/l (milliequivalents per liter) Ca: meq/l Mg = 5:1" in Figure 17 represents the most common magnesium-calcium relation of limestone aquifers. The magnesium in low-magnesium calcite is less than ten percent of the molal weight of the calcium of the calcite. The magnesium in high-magnesium calcite is greater than ten percent of the molal weight of the calcium of the

70
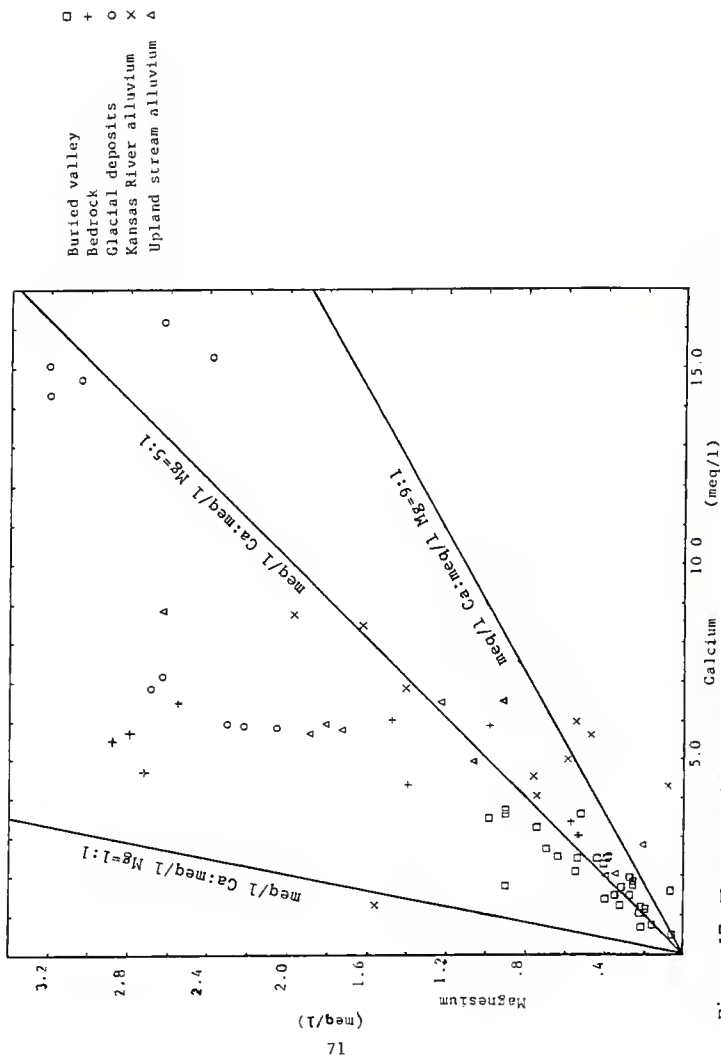
Figure 17. The magnesium-calcium relationships of the waters in the area of investigation.

calcite. The line labeled meq/1 Ca: meq/1 Mg = 9:1 represents the
boundary between low- and high-magnesium calcites. The magnesium-cal-
cium relation that waters dissolving dolomite should have is repre-
sented by the line labeled "meq/1 Ca: meq/1 Mg = 1:1" (Erikson, 1985;
Meisler and Bechler, 1967). The magnesium-calcium relation of the
sample waters of the area ranges from approximately 1.10 to 0.0 (Fig.
17).

Water samples with calcium concentrations up to between 6.0 and
8.0 meq/1 tend to occur close to the 5:1 calcium-magnesium line (Fig.
17). Above calcium concentrations of 6.0 meq/1 the magnesium-calcium
relations follow a nearly vertical trend where the magnesium concen-
tration doubles and the calcium concentration increases only slightly.
The small change of the calcium concentrations above 6.0 meq/1 and the
great change of the magnesium concentrations is probably caused by the
sample waters becoming saturated with respect to calcite, but conti-
nuing the dissolution of dolomite and gypsum (Bricker and Garrels,
1968).

The water samples from the glacial deposits have the largest
calcium and magnesium concentrations of the hydrologic groups (Table
8; Fig. 17). The water samples of the glacial deposits have high mag-
nesium concentrations of ranging up to 4.0 meq/1, but the magnesium
concentrations are low relative to calcium concentrations that often
are greater than 14.0 meq/1 calcium. The low magnesium-calcium rela-
tion of the glacial deposits is probably due to the dissolution of
gypsum, or recharge containing dissolved gypsum from the bedrock, add-
ing calcium and very little magnesium to the water. The presence of

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Table 8. The average, low, and high concentrations of the chemical constituents of the hydrologic units in southwest Pottawatomie County, Kansas.

| | KANSAS RIVER A. | | | | GLACIAL DEPOSITS | | | | UPLAND STREAM A. | | | | BEDROCK | | | | BURIED VALLEY SED. | | | |
	MEAN	MED.	LOW	HIGH	MEAN	MED.	LOW	HIGH	MEAN	MED.	LOW	HIGH	MEAN	MED.	LOW	HIGH	MEAN	MED.	LOW	HIGH
TEMP.	15	14.6	13.1	18.7	15.1	14.4	11	18.9	15.4	15.1	11.8	20.1	14.6	14.4	11.3	19.6	14.8	14.8	7.3	19.5
SP. COND.	929	882	596	1434	1837	1597	973	3948	2252	552	282	9267	896	931	386	1478	363	370	86	646
pH	7.2	7.2	7.0	7.4	6.8	6.8	6.2	7.3	6.9	7.1	6.5	7.2	7.1	7.1	6.9	7.5	7.0	7.0	6.4	7.6
CALCIUM	124	117	81	175	215	215	117	324	102	115	40	177	96	101	61	130	40	36	10	74
MAGNESIUM	13	9.7	5.8	24	32	32	25	39	15	14	2.6	32	20	18	6.6	35	5.2	4.4	12	.70
SODIUM	25	26	12	35	62	59	14	140	27	26	1.6	64	35	13	6.5	85	21	20	9.0	37
POTASSIUM	13	10	3.2	28	2.5	2.5	1.1	4.4	2.1	2.1	.80	5.1	1.8	1.3	.8	3.7	2.1	2.2	.70	3.3
STRONTIUM	.47	.45	.26	.78	.86	.46	DL	3.8	.83	.90	.14	1.8	8.5	1.4	DL	26	.25	.18	DL	.57
IRON	1.31	.05	DL	4.5	.63	.13	DL	5.1	3.2	1.5	DL	13.8	.07	.05	DL	DL	.10	DL	DL	.63
MANGANESE	.125	DL	DL	.70	.22	.05	DL	2.4	.37	.09	DL	1.3	.04	DL	DL	.2	.13	DL	DL	.75
BICARB.	343	328	291	374	378	376	239	511	358	400	138	455	341	369	188	457	179	161	56	292
SULFATE	77	61	10	189	348	226	44	970	49	31	3.0	247	122	31	DL	519	11	8.5	DL	92
CHLORIDE	24	17	6	56	49	43	12	103	18	13	7.0	65	6.0	11	2.0	38	5.8	5.0	1.0	14
NITRATE	49	34	1.0	220	71	12	7.0	192	9.0	8.0	6.0	13	18	12	8.0	54	23	10	DL	68
NO OF ANALYSES	10 (5)				10 (6)				10 (5)				10 (6)				30 (18)			

Note: Concentrations are expressed as milligrams per milliliter. Note: Specific conductivities are expressed as micromhos per centimeter squared. Note: The number in the parenthesis is for nitrate analyses.

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dissolved gypsum in water would also explain the water samples from the Kansas River alluvium that have magnesium-calcium relations falling below the 1:9 line (Fig. 17).

The water samples from the buried valley sediments have the lowest calcium and magnesium concentrations of the hydrologic units (Table 8). The water samples from the buried valley sediments form a small group of points with very little scatter that occur slightly above the 5:1 calcium-magnesium line (Fig. 17). This seems to indicate that most of the calcium and magnesium of the buried valley sediments came from the dissolution of the typical limestone of the area.

Potassium-Sodium Ratio--The potassium-sodium relations of sample waters from hydrologic units other than the buried valley show two distinctive groups, samples with high potassium concentrations and samples with high sodium concentrations (Fig. 18).

Water samples containing high sodium concentrations occur in the glacial deposits and bedrock and have potassium concentrations that are less than .15 meq/l (Fig. 18). The sodium concentrations of these samples range from 0.0 to 6.2 meq/l and many are at least an order of magnitude greater than the corresponding potassium concentration.

The lowest potassium concentrations occur in water samples from the bedrock and are due to sodium-rich illite being the dominant clay mineral in the bedrock (Table 8; Fig. 18; Lloyd and Heathcote, 1985; Matthess, 1982; Twiss, 1955; Hargedine, 1963; Dowling, 1967; Dulekoz, 1966; Watkins, 1957; Schmidt, 1974). The illite of the bedrock con-

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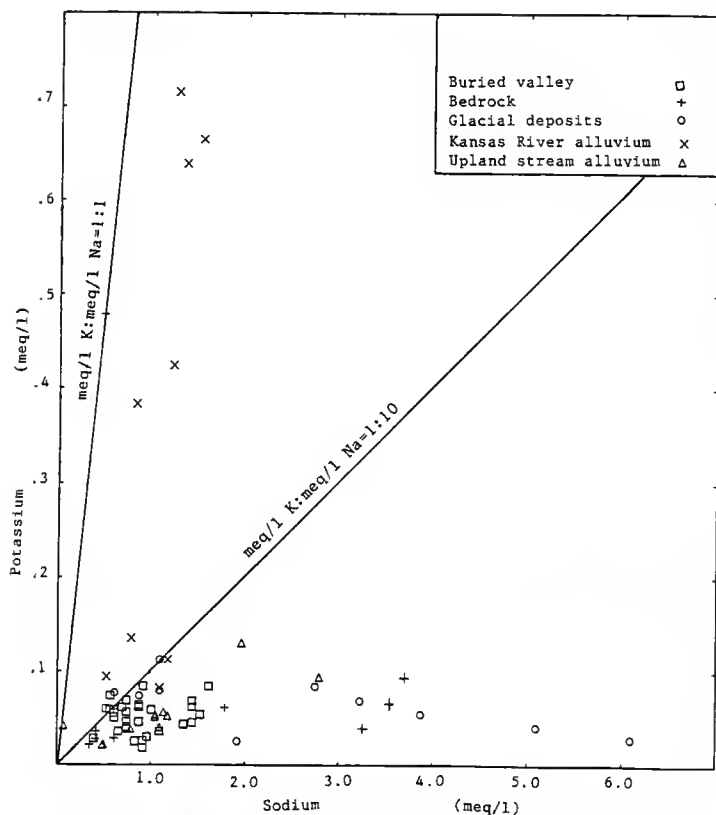
Figure 18. The potassium-sodium relationships of the waters in the area of investigation.

tains large amounts of sodium because it was exposed to seawater. Sodium has a weak attraction to the surface of clay particles, and it is exchanged for calcium, magnesium, and potassium from groundwater (Lloyd and Heathcote, 1985; Erikson, 1985). The calcium and magnesium ions taken out of solution are replaced by ions dissolved from carbonates (Erikson, 1985; Lloyd and Heathcote, 1985). The sodium concentrations of water contacting shale in the bedrock increase, and the potassium concentrations decrease (Hem, 1970).

The glacial deposits were deposited by fresh water and wind, and cation exchange with clay minerals cannot account for the high sodium concentrations of water from the glacial deposits (Scott and others, 1959; Lloyd and Heathcote, 1985). Two alternate sources of sodium in the glacial deposits are recharge by water from the bedrock, and pollution by animal wastes and sewage. Many of the water samples containing high sodium concentrations also contain the high concentrations of nitrate and chloride that are characteristic of pollution by animal wastes and sewage (Hem, 1970; Matthess, 1982).

Water samples from the Kansas River alluvium have the highest concentrations of potassium in the area (Table 8; Fig. 18). The sodium concentrations of the Kansas River alluvium are low and remain under 2.0 meq/l. The Kansas River valley is the most extensively cultivated portion of the area, and many of the water samples containing high potassium also contain high nitrate indicating that the most probable source of potassium in the Kansas River alluvium are fertilizers used on the crops of the Kansas River Valley.

The water samples from the buried valley sediments are low in

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both sodium and potassium, and form a small group with very little scatter in the points (Table 8; Figure 18).

Calcium-Bicarbonate Ratio--The ratio of the equivalents per liter of calcium and bicarbonate produced by the dissolution of calcite is represented by the line marked meq/1 Ca: meq/1 HCO $_3$ = 1:1 in Figure 19 (Table 6). The dissolution of dolomite produces 2.0 meq/1 bicarbonate for every 1.0 meq/1 calcium and is represented by the line marked meq/1 HCO$_3$: meq/1 Ca = 2:1. The dissolution of gypsum will make the calcium concentrations higher than the bicarbonate concentrations and, a line labeled meq/1 HCO$_3$: meq/1 Ca = 1:2 is shown as a reference. Most of the samples for the investigation are located between the 1:1 and 2:1 bicarbonate-calcium relations. Bicarbonate may also come from the dissolution of carbon dioxide gas from the atmosphere and the decay of organic materials.

Water samples from the glacial deposits have the largest bicarbonate concentrations of the hydrologic units, yet the bicarbonate-calcium ratios of the water samples are low with four of the ten water samples having bicarbonate-calcium ratios less than 1:2 (Table 8; Fig. 19). Many of the water samples from the Kansas River alluvium contain bicarbonate-calcium ratios less than 1:1. The high calcium concentrations of these water samples may be caused by the dissolution of gypsum, or recharge by gypsum-rich waters. Dissolved gypsum can make the bicarbonate concentration of water decrease by increasing the calcium concentration of the water until calcite precipitates.

The bicarbonate-calcium ratios of water samples from the upland

77
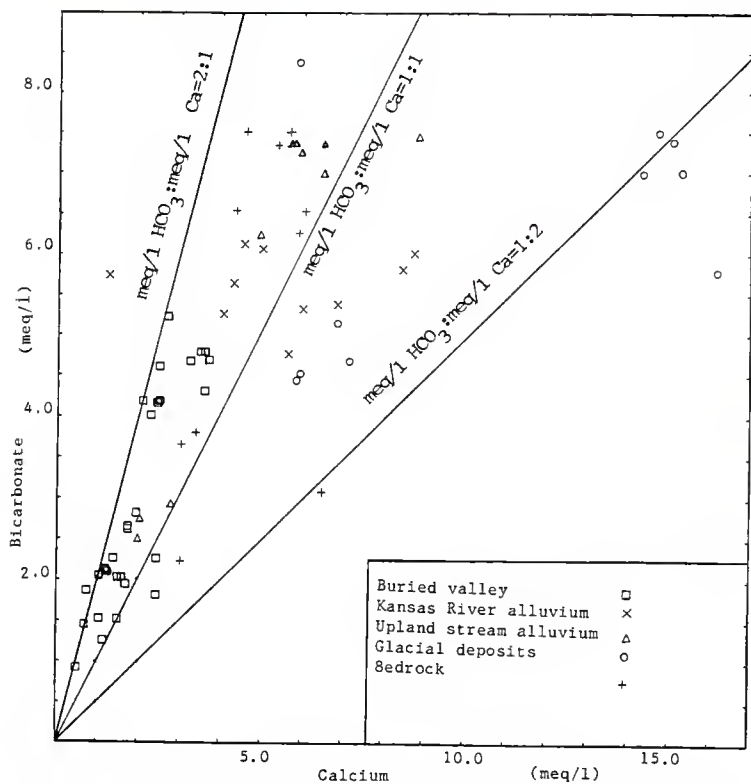
Figure 19.   The calcium-bicarbonate relationships of the waters in the
area of investigation.

stream alluvium vary a great deal (Fig. 19) probably because the cal-
cium concentrations of the sources of recharge for the upland stream
deposits vary greatly. The upland stream alluvium might receive water
from the bedrock or glacial deposits and contain high calcium concen-
trations, or it might receive recharge from the buried valley sedi-
ments and contain low calcium concentrations.

    Water samples from the buried valley sediments have both the
lowest bicarbonate concentrations, and bicarbonate-calcium relations
of the hydrologic units (Table 8; Fig. 19). The small amount of cal-
cium in the water samples from the buried valley sediments suggests
that most of the bicarbonate in solution came from the atmosphere and
soil (Snow, 1963).

Sulfate-Calcium Ratio--The dissolution of gypsum produces equal equi-
valents per liter of sulfate and calcium, and it is represented by the
line marked "meq/1 Ca:meq/1 SO_4 = 1:1" in Figure 20 (Table 6). The
dissolution of many minerals affect the calcium concentration of
water.

    Calcium concentrations are higher than the sulfate concentrations
in all but three of the sample waters (Fig. 20). The high calcium and
low sulfate concentrations of the sample waters indicate that the dis-
solution of calcite and dolomite control the calcium concentrations of
most water in the the area.

    The three samples with calcium-sulfate ratios higher than 1:1
come from three different hydrologic units (Fig. 20). There are se-
veral ways water may develop high sulfate concentrations. If the cal-
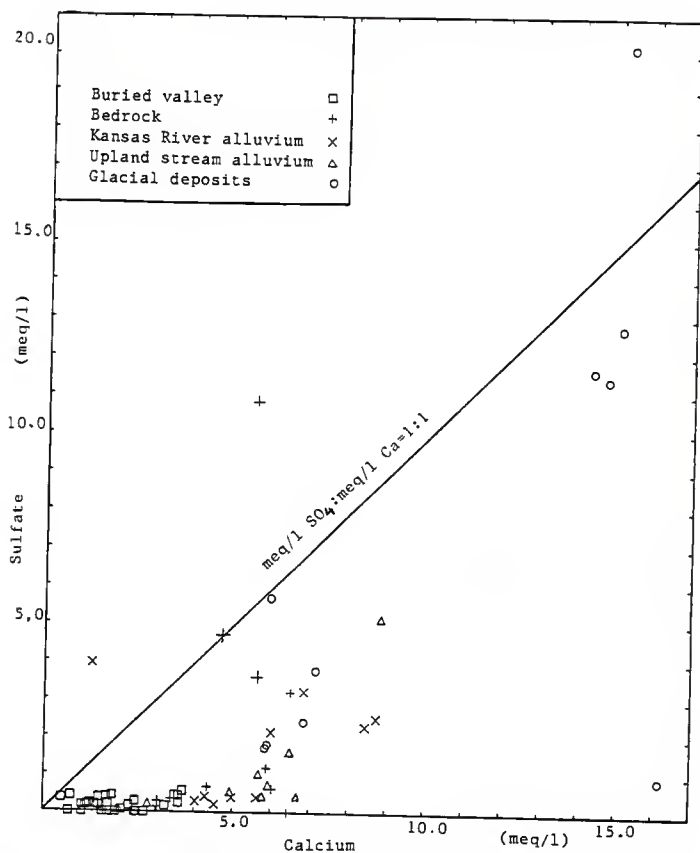
Figure 20. The calcium-sulfate relationships of the waters in the area of investigation.

cium concentration of water increases to the point where calcite will precipitate, then the sulfate-calcium ratio will increase (Hem, 1970; Lloyd and Heathcote, 1985). Also, calcium may be removed from solution by cation exchange with clay minerals (Lloyd and Heathcote, 1985). Finally, sulfate might be increased in water by the dissolution of celestite or sulfide minerals in bedrock (Hem, 1970; O'Connor, 1971).

The water samples containing the greatest sulfate concentrations came from the glacial deposits (Fig. 20; Table 8). The high sulfate concentrations of the glacial deposits indicate that the glacial deposits may be recharged from the neighboring bedrock (Erikson, 1985; Lloyd and Heathcote, 1985). Gypsum may, on rare occasions, be deposited in glacial sediments, but the small thicknesses of the glacial sediments in the area make it unlikely that solid gypsum could exist in them.

Water samples from the buried valley sediments have uniformly low sulfate concentrations that never exceed 1.0 meq/l (Table 8; Fig. 20). The small sulfate concentrations in water samples from the buried valley sediments suggests that the amount of recharge to the buried valley sediments from other hydrologic units is small. The sulfate concentrations of water from the buried valley sediments would be higher if significant amounts of recharge were received from other hydrologic units.

Chloride-Sodium Ratio--The dissolution of solid sodium chloride produces equal equivalents per liter of sodium and chloride, a ratio represented by the line labeled "meq/l Cl: meq/l Na = 1:1" in Figure 21.

81
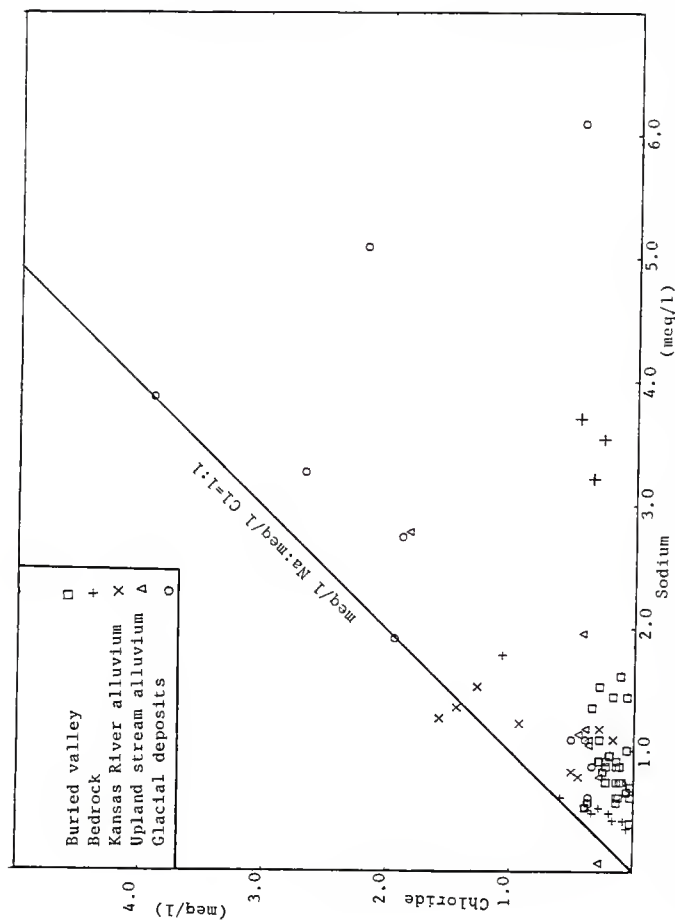
Figure 21. The sodium- chloride relationships of the waters in the area of investigation.

The chloride-sodium relations of the sample waters, with the exception of samples from the buried valley sediments, are extremely variable (Fig. 21; Table 8). The chloride-sodium ratios of most sample waters are less than 1:1, indicating that sources of sodium with low centrations are common in the area.

Water samples from the bedrock contain the lowest chloride concentrations of the hydrologic units (Table 8; Fig. 21). Some water samples from the bedrock contain high sodium and low chloride. The most common source of sodium without chloride is ion exchange, where clay minerals exchange sodium for calcium, magnesium, and potassium. Sodium may come from rocks containing gypsum, but sodium from the gypsum deposits should contain chloride.

The chloride-sodium relations of water samples from the glacial deposits and Kansas River alluvium are widely scattered (Fig. 21). The water samples from the glacial deposits contain the highest concentrations of chloride and sodium, but water samples from the Kansas River alluvium contain high concentrations of sodium and chloride as well. Many samples containing high concentrations of chloride, sodium and nitrate come from wells located close to animal pens, and appear to have been polluted by animal wastes.

Water samples from the buried valley sediments have low chloride concentrations (Table 8; Fig. 21). Some areas of the buried valley have many homes built in them, but chloride from pollution is low, possibly due to the watertable occurring at great depths. A possible source of sodium, calcium, and potassium in the waters of the buried valley is the weathering of the plagioclase of the buried valley

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sediments deposits (Snow, 1963).

Nitrate

Nitrate is an ion which is harmful to humans in large quantities, and water containing large nitrate concentrations may contain harmful bacteria because nitrate is a food source for bacteria (Hem, 1970; Kansas Water Resources Board, 1959). Large concentrations of chloride, sodium, and nitrate are often found where pollution by animal wastes and sewage has occurred, and large amounts of potassium, phosphate, and, occasionally, calcium occur where nitrate fertilizers are used. Large concentrations of nitrate and small concentrations of bicarbonate are found where nitrate pollution from decaying vegetation has occurred.

Nitrate pollution is a common problem in northeast Kansas (Denne, 1980; Kansas Water Resources Board, 1959). The nitrate concentrations of Kansas are generally less than 10 mg/l, but the nitrate concentrations of many wells in northeast Kansas are above the health limit of 45 mg/l nitrate (Denne, 1980; Kansas Water Resources Board, 1959). Nine of the 38 samples tested for nitrate contained water with nitrate greater than 45 mg/l (Appendix III). Large concentrations of chloride, sodium, and nitrate occurred together in the water of 4 of these samples, and large concentrations of potassium and nitrate occurred together in the water of 4 of the remaining samples. Small concentrations of bicarbonate and large concentrations of nitrate occurred in one sample.

Large amounts of nitrate are often associated with shallow wells

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because the nitrate cannot spread out over a large area and be diluted in a shallow aquifer, and there is less opportunity for bacteria to breakdown the nitrate. Deep wells occur in the bedrock and buried valley sediments and water from both hydrologic units had low nitrate values. The most highly developed portions of the area, the glacial deposits and the Kansas River alluvium, had the highest nitrate concentrations of the investigation.

Geographical Variation of the Chemical Constituents

Analyses of the variance of the ion concentrations and the saturation indices of calcite, dolomite, and gypsum between the water samples from the buried valley sediments and the water samples that were not from the buried valley sediments were performed using the minitab statistical program (Appendix X; Ryan and others, 1985). The isocon maps of the ion concentrations were produced using the surface II contouring program (Sampson, 1975).

The result of an analysis of variance (calculated F value) is calculated by using the equations (Johnson, 1980),

$$F = \frac{MS(factor)}{MS(error)} \tag{17}$$

$$MS(factor) = \frac{\Sigma (Ti^2)/C - (\Sigma x^2)/n}{df(factor)} \tag{18}$$

85

$$MS(error) = \frac{\Sigma(X^2) - \Sigma(Ti^2)/C}{df(error)} \quad (19)$$

where, MS(factor) = mean square of the factor between

the sample groups

MS(error) = mean square of error within the sample groups

df(factor) = number of sample groups – 1

df(error) = sum of all the samples–the number of

sample groups

Ti = sum of the values of the samples of a group

C = the number of replicated values

n = the number of samples

X = the values of the samples

The calculated F value is compared to values of the F distribution that is defined by the degrees of freedom and the significance level (the probability of rejecting a true hypothesis) desired for the analysis of variance (Johnson, 1980). If the calculated value of F is greater than the value of F listed for a significance level, then the sample groups come from different populations. The analyses of variance for the saturation indices and all the ions, except iron and strontium indicate that there is better than a 95 percent chance that the water samples from the buried valley sediments are different from the water samples from the other hydrologic units (Table 9).

The water samples containing the largest calcium, magnesium, bicarbonate, and sulfate concentrations occur in the glacial deposits

Table 9. The results of the analyses of variance between the sample from the buried valley and the samples from the other hydrologic units.

Degrees of Freedom
 Factor = 1
 Error = 68

Number of Samples
 Buried Valley Sediments 30
 Other Hydrologic Units   40

At a Confidence Level of .05 F = 4.10
At a Confidence Level of .025 F = 5.45
At a Confidence Level of .01 F = 7.36

Characteristic or Ion Tested	Calculated F Value
Calcium	59.40
Magnesium	48.59
Sodium	7.54
Potassium	4.94
Strontium	3.96
Iron	5.08
Manganese	2.39
Bicarbonate	67.35
Sulfate	11.84
Chloride	17.70
Saturation Index of Calcite	29.25
Saturation Index of Dolomite	46.00
Saturation Index of Gypsum	10.73

and bedrock of the central and eastern portions of the area (Fig. 22, 23, 24, and 25). The bedrock of these areas is highly dissected by stream valleys, and many wells that obtain water from the bedrock occur large distances away from the area of recharge. Water flowing long distances through the bedrock has long periods of time and the opportunity to dissolve large amounts of soluble minerals (Erikson, 1985; Lloyd and Heathcote, 1985). The ion concentrations of some water samples from the glacial deposits in the northern and central portions of the area are greater than those of any sample from the bedrock. The small grain size of the glacial deposits probably causes recharge from the bedrock to occur slowly allowing the water time to dissolve soluble minerals from the bedrock (Walters, 1953; Strausberg, 1972).

Surprisingly, the glacial deposits and bedrock of the western portion of the area contain water with uniformly low calcium, magnesium, bicarbonate, and sulfate concentrations (Fig. 22, 23, 24, and 25). Only the bicarbonate concentrations vary enough to define the western and eastern bedrock highs of the area. The low ion concentrations of this area are probably due to the structure of the bedrock.

The bedrock in the western portion of the area is recharged at outcrops on the eastern flank of the Flint Hills and the water flows westward, into the Flint Hills. Thus, wells obtaining water from the bedrock in the western portion of the area probably obtain water that has not had the opportunity, or time, to dissolve many minerals.

The water samples with the lowest calcium, magnesium, bicarbonate, and sulfate concentrations occur in the buried valley sediments, follow the trend of the buried valley, and define the area of the buried

88
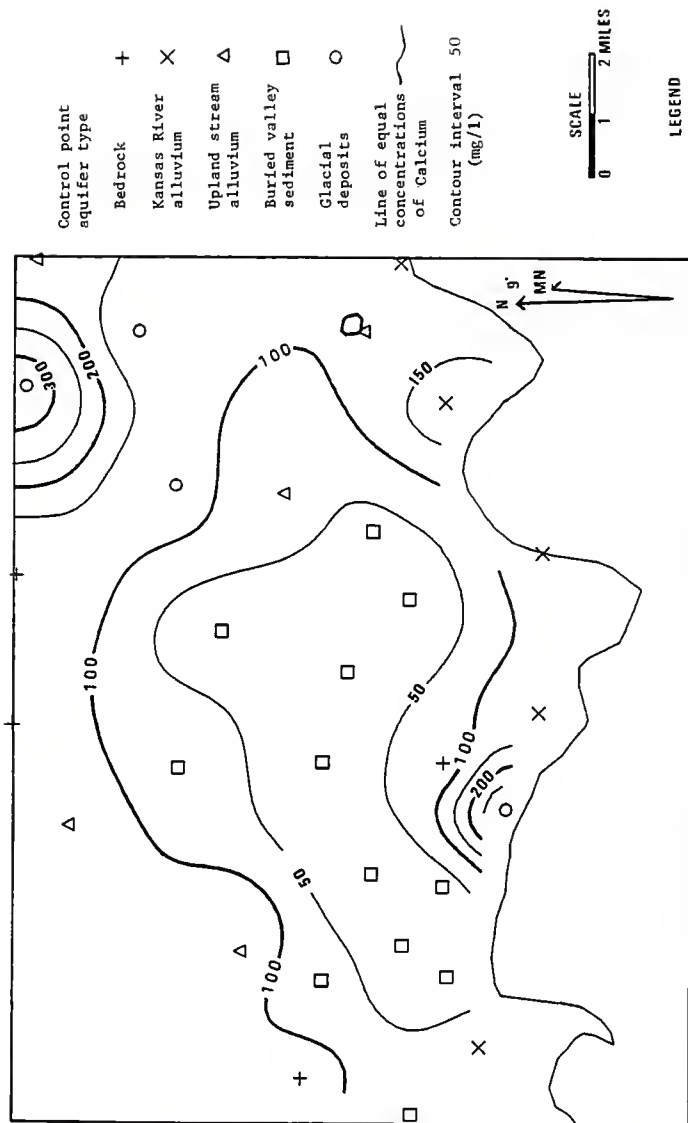
Figure 22. Calcium concentration of the groundwater (mg/l) in southwest Pottawatomie County, Kansas for the first series of samples.

LEGEND

Control point
aquifer type

Bedrock   +

Kansas River
alluvium   ×

Upland stream
alluvium   △

Buried valley
sediment   □

Glacial
deposits   ○

Line of equal
concentrations
of Calcium

Contour interval 50
(mg/l)

SCALE

0   1   2 MILES
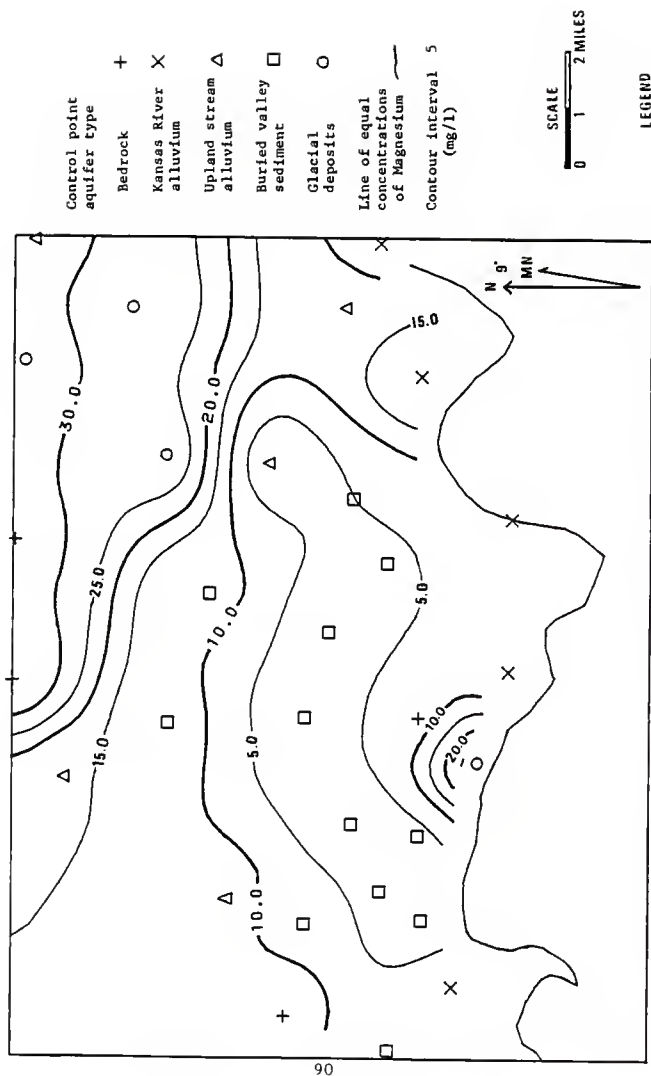
Figure 23. Magnesium concentration of the groundwater (mg/l) in southwest Pottawatomie County, Kansas for the first series of samples.
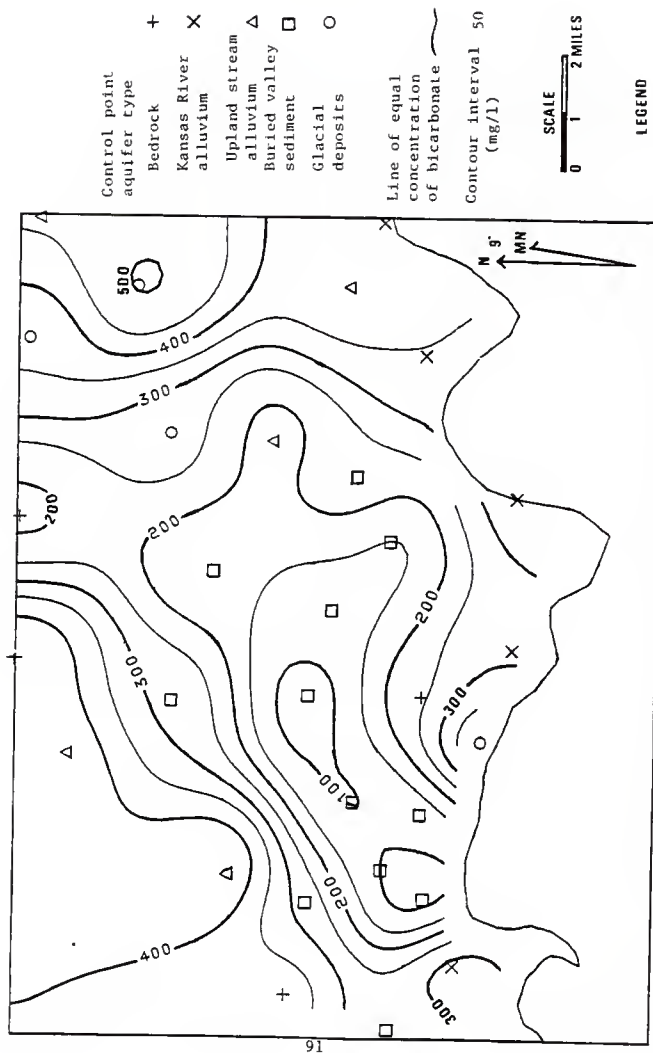
Figure 24. Bicarbonate concentration of the groundwater (mg/l) in southwest Pottawatomie County, Kansas for the first series of samples.

LEGEND

Control point
aquifer type

Bedrock   +

Kansas River   ✗
alluvium

Upland stream   △
alluvium

Buried valley   ◻
sediment

Glacial   ○
deposits

Line of equal   ⌒
concentration
of bicarbonate

Contour interval   50
(mg/l)

SCALE

0   1   2 MILES

LEGEND

Control point aquifer type	
Bedrock	+
Kansas River alluvium	✕
Upland stream alluvium	△
Buried valley sediment	☐
Glacial deposits	○

Line of equal concentrations of Sulfate

Contour interval 100 (mg/l)

SCALE

0   1   2 MILES
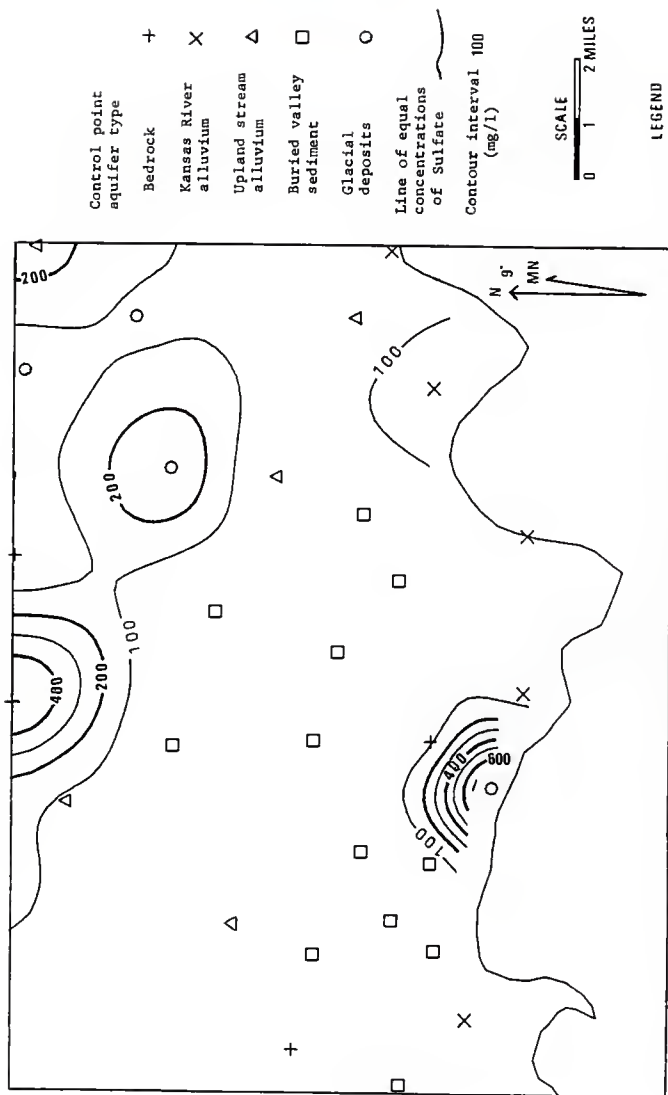
Figure 25. Sulfate Concentration of groundwater (mg/l) in southwest Pottawatomie County, Kansas for the first series of samples.

valley (Fig. 22, 23, 24, and 25). The low ion concentrations of water from the buried valley sediments are probably due to two factors. The first is the small amount of soluble minerals in the buried valley, where the sediments consist of chiefly of quartz, plagioclase and illite (Snow, 1963). The second factor seems to be the lack of significant recharge of the buried valley sediments by water containing large concentrations of ions.

Geographically, the potassium concentrations of the area are nearly constant, and are close to 2.0 mg/l over most of the area (Fig. 26). The highest potassium concentrations occur in the Kansas River and lower Vermillion Creek valleys, and are believed to be due to pollution by potassium-rich fertilizers.

Geographically, the highest sodium and chloride concentrations occur in water samples from the glacial deposits and Kansas River alluvium of the eastern portion of the area (Fig. 27; 28). The high sodium and chloride concentrations of these areas are often associated with high nitrate concentrations indicating that pollution due to animal wastes is a likely source of the sodium and chloride (Denne, 1980; Matthess, 1982). The water samples with the lowest sodium and chloride concentrations come from the bedrock and buried valley sediments.

Figure 29 is an isocon map of the iron concentrations of the area. The geographic variations of the manganese concentrations of the sample waters are similar to those of the iron concentrations.

Water samples from the Kansas River alluvium, upland stream alluvium, and glacial deposits of the eastern and central portions of the area contain the highest iron concentrations (Fig. 29). The high iron

Figure 26. Pottassium concentration of the groundwater (mg/1) in southwest Pottawatomie County, Kansas for the first series of samples.

LEGEND

Control point aquifer type

Bedrock   +

Kansas River
alluvium   ×

Upland stream
alluvium   △

Buried valley
sediment   □

Glacial
deposits   ○

Line of equal
concentrations
of Potassium

Contour interval .2
(mg/1)

SCALE

0   1   2 MILES
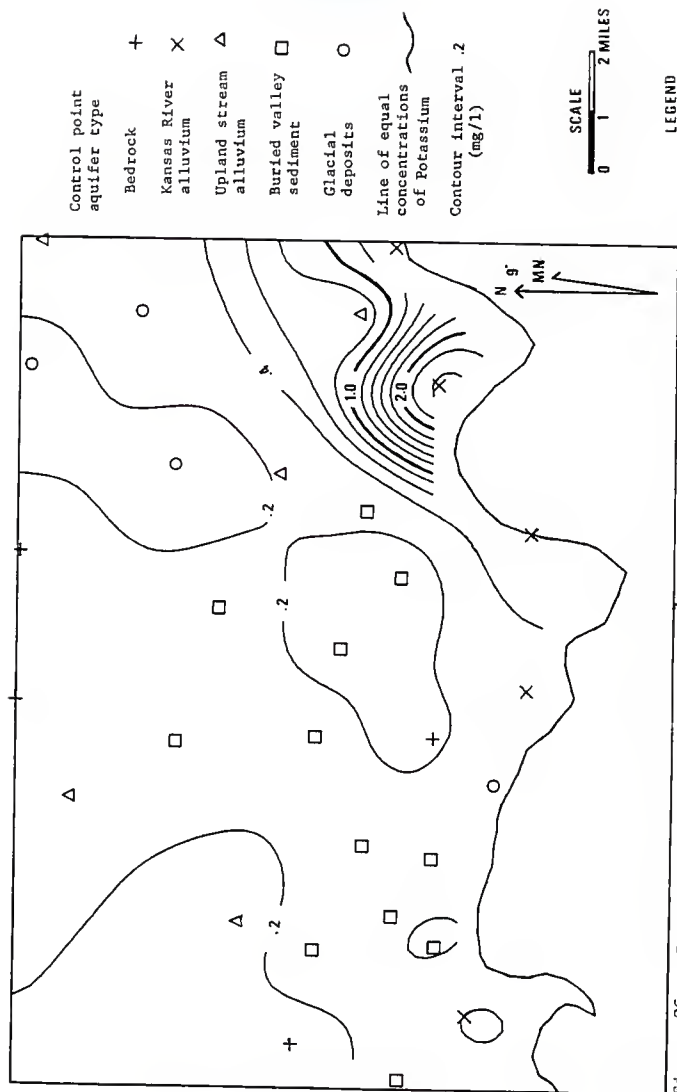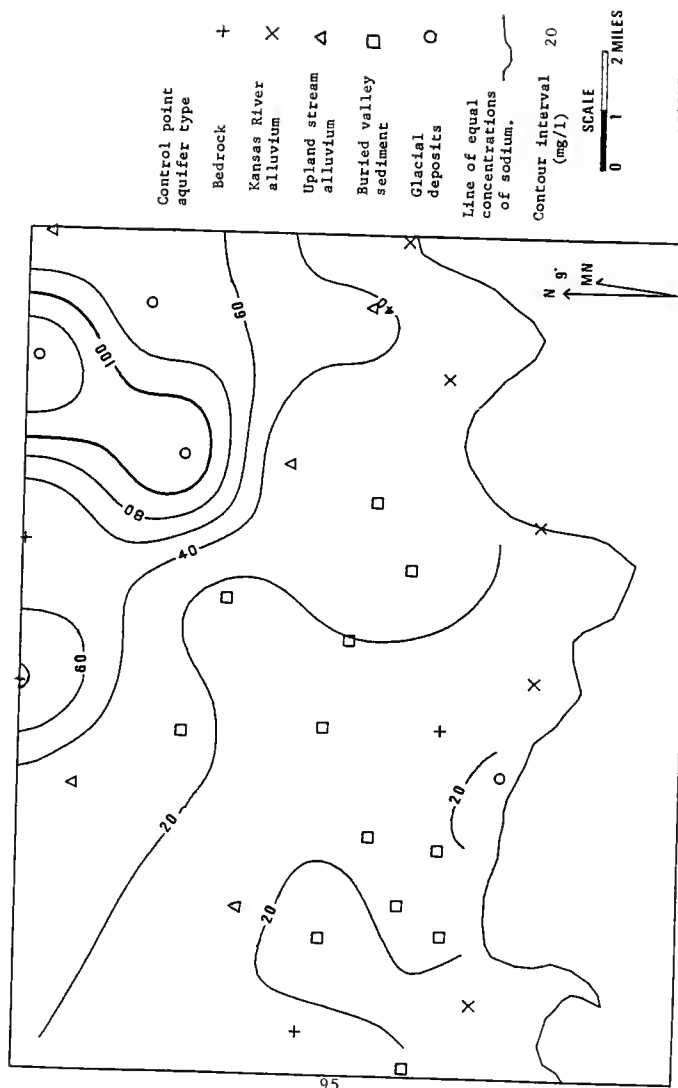
Figure 27. Sodium concentration of the groundwater (mg/l) in southwest Pottawatomie County, Kansas for the first series of samples.
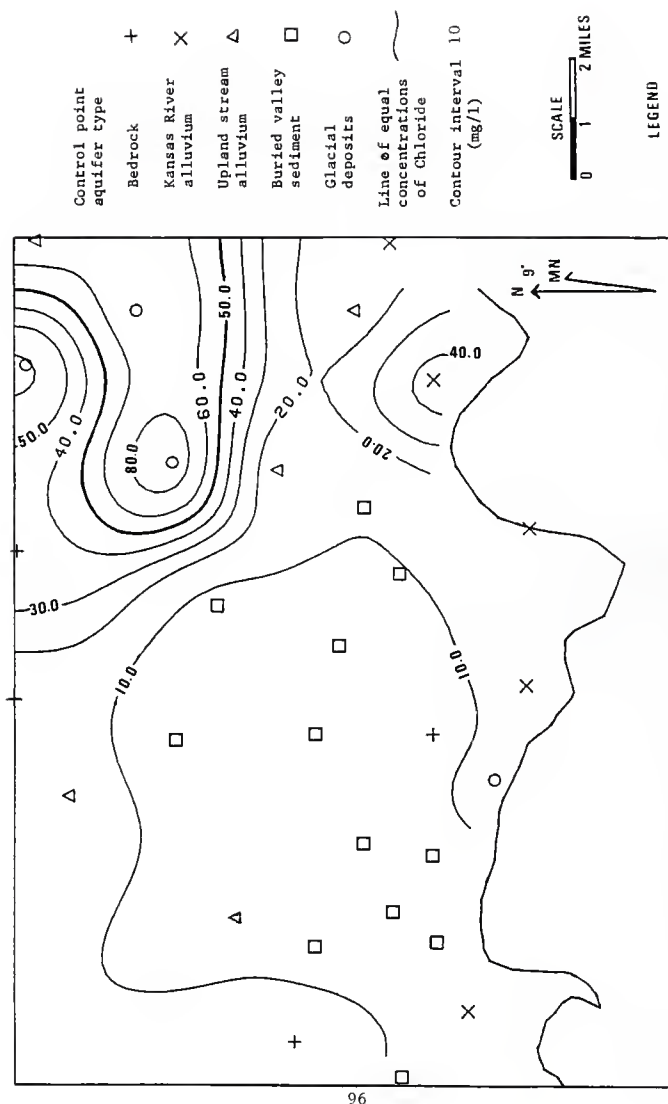
Figure 28. Chloride concentration of groundwater (mg/1) in southwest Pottawatomie County, Kansas for the first series of samples.

LEGEND

Control point aquifer type

Bedrock $+$

Kansas River alluvium $\times$

Upland stream alluvium $\triangle$

Buried valley sediment $\square$

Glacial deposits $\bigcirc$

Line of equal concentrations of Chloride

Contour interval 10 (mg/1)

SCALE

0   1   2 MILES
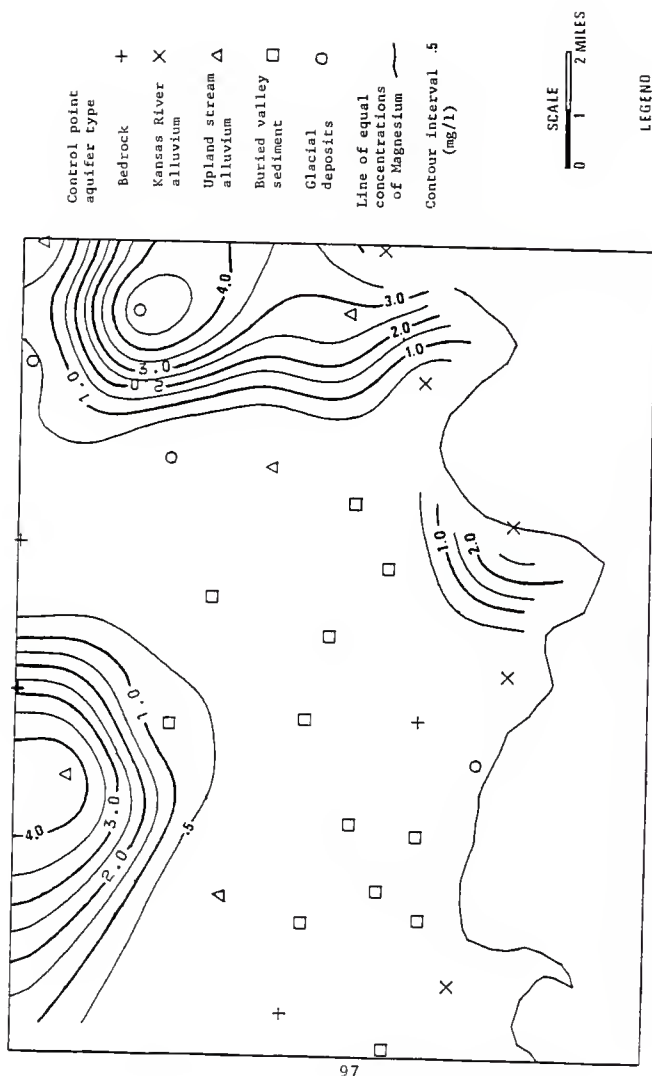
Figure 29. Iron concentration of the groundwater (mg/1) in southwest Pottawatomie County, Kansas for the first series of samples.

LEGEND

Control point
aquifer type

Bedrock   +

Kansas River   ✗
alluvium

Upland stream   △
alluvium

Buried valley   □
sediment

Glacial   ○
deposits

Line of equal   ⌒
concentrations
of Magnesium

Contour interval .5
        (mg/1)

SCALE

0   1   2 MILES

97

concentrations of the glacial deposits, Kansas River alluvium, and up-
land stream alluvium may be due to the soils contain large amounts of
organic materials that decay and form carbonic acid. The acid dis-
solves iron which is abundant in the glacial deposits, Kansas River
Alluvium, and the upland stream alluvium, and carries it to the water
table (Hem, 1970; Whittemore, 1976). The high iron concentrations may
also occur where bacteria reduce sulfate to the S^{2-} ion and use the
oxygen of the sulfate to digest organic material. The sulfur ion re-
acts with hydrogen to form hydrogen sulfide, and the water becomes a-
cidic. The acidic water then dissolves large amounts of iron and man-
ganese (Hem, 1970; Lloyd and Heathcote, 1985; Matthess, 1982).

The lowest concentrations of iron occur in water samples from the
buried valley and the bedrock west of the buried valley (Fig. 29). The
low iron concentrations of water samples from the buried valley are
probably due to the lack of iron and manganese minerals in the sedi-
ments of the buried valley (Snow, 1963). The low iron and manganese
concentrations in the water samples from the bedrock are probably due
to the small amount of time the water has been in the aquifer.

The well with a concentration of 26 mg/l strontium was not used
as a data point for the isocon map of strontium because the isopach
contour interval that would be needed to show 26 mg/l strontium would
not adequately show the strontium variations of the other wells (Fig.
30). Strontium concentrations are generally highest in water samples
from the bedrock and glacial deposits of the western portion of the
area. The high strontium concentrations in water samples from the bed-
rock and the glacial deposits are probably the result of the dissolu-
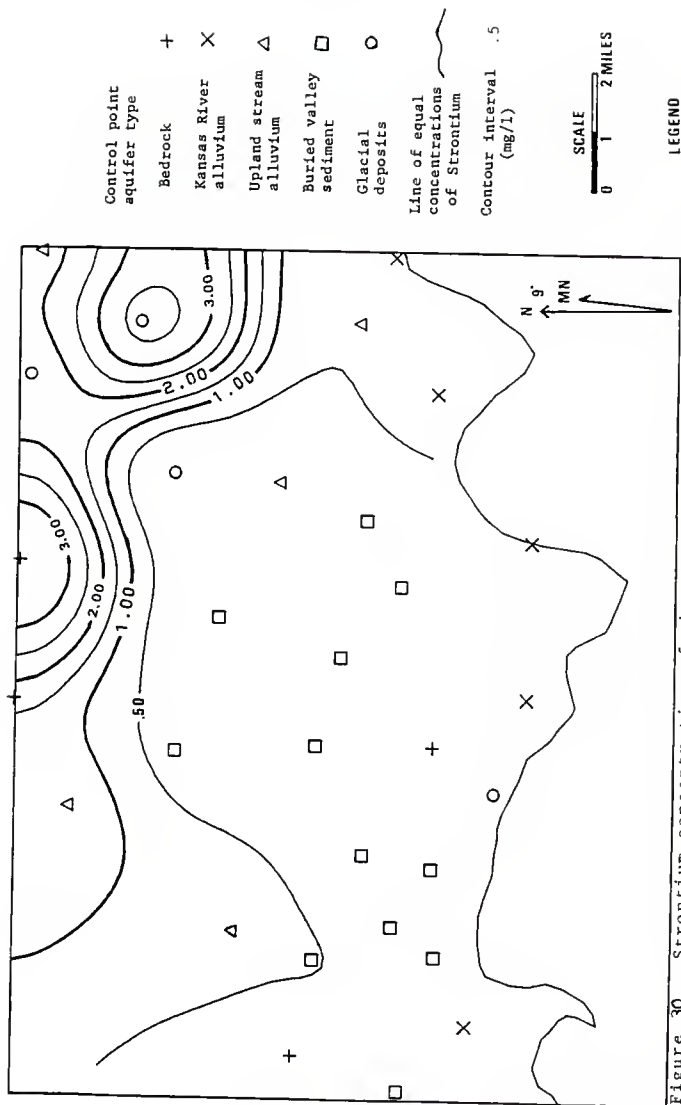
Figure 30. Strontium concentration of the groundwater (mg/1) in southwest Pottawatomie County, Kansas for the first series of samples.

Control point
aquifer type

Bedrock   +

Kansas River
alluvium   ✕

Upland stream
alluvium   △

Buried valley
sediment   ▢

Glacial
deposits   ○

Line of equal
concentrations
of Strontium

Contour interval   .5
(mg/1)

SCALE

0   1   2 MILES

LEGEND

tion of celestite in the bedrock. Strontium occurs in calcite by replacing the calcium ion in the crystal structure, but calcite dissolution will add only small amounts of strontium to water (Hem, 1970; Krauskopf, 1979). The strontium concentrations of 26 mg/l occurred in the well of a cattle ranch and may come from the residue of strontium-rich cattle feed (Chaudhuri, 1986, personal communication).

The strontium concentrations of water samples from the buried valley sediments are the lowest of the hydrologic units, and the strontium is detectable only in the most mineralized water samples (Fig. 30).

Seasonal Variations

Changes in chemical concentrations with time occur because of seasonal changes of precipitation and the activity of man. The simplest temporal variation is caused by precipitation infiltrating into an aquifer, but the causes of temporal variations may be much more complex than dilution by infiltration.

The concentration changes of the waters of the five hydrologic units with season were evaluated using the chemical data from the 16 wells sampled for the first second and third sample series and the eight samples from the fourth sample series. The average concentrations of each series of analyses for a hydrologic unit (Appendix III) and the monthly precipitation occurring at Wamego, Ks., were used to plot Figures 31, 32, 33, 34, and 35 (National Atmospheric and Oceanic Administration, 1985; 1986).

Buried Valley Sediments--The changes of the water quality with time of the buried valley sediments are the smallest in the area. The small concentration changes may be caused by the water from the buried valley sediments containing such low ion concentrations that it does not differ greatly in quality from water infiltrating into the buried valley sediments (Fig. 31; Whittemore and Switek, 1977).

The ions in the water from the buried valley sediments having the greatest relative concentration change with time are sulfate and chloride, and the ions with the smallest relative concentration change with time are calcium and magnesium (Fig. 31). The bicarbonate concentration changes mirror those of calcium and magnesium, and indicate that water recharging the buried valley sediments contains some dissolved calcium carbonate, but the solubility limit of calcite is not reached. The change of the sulfate and chloride are probably caused by sulfate-rich recharge being diluted by varying amounts of sulfate-poor water after entering the aquifer.

Calcium, magnesium, chloride, potassium, and sulfate have their smallest concentrations in the sample series 3 months after the period of greatest precipitation (Fig. 31). These concentration changes suggest that effects of precipitation reach the buried valley sediments within three months. The potassium concentrations are greatest during the spring, when fertilizer is often used on crops, and the bicarbonate concentrations are greatest during the winter.

Upland Stream and Kansas River Alluvium--The ions in the water from the upland stream alluvium having the greatest relative concentration

101
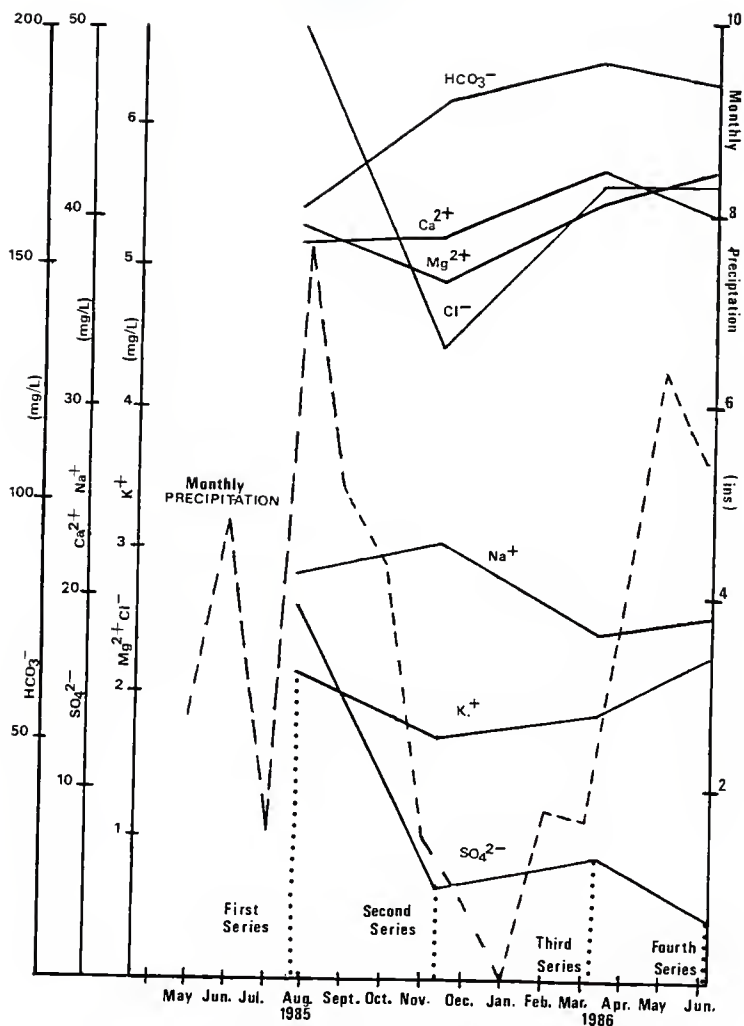
Figure 31. The change in precipitation and in the average composition of water from the buried valley sediments during the period of investigation.

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change with time are sodium, chloride, and sulfate (Fig. 32; 33). The
ions in the water from the Kansas River alluvium that have the great-
est relative concentration change with time are sulfate and potassium.
The potassium concentrations in water from the Kansas River alluvium
are greatest during the spring and summer suggesting that the potas-
sium comes from fertilizer. The great change of the sulfate is proba-
bly caused by sulfate-rich recharge being diluted by varying amounts
of sulfate-poor water after entering the aquifer.

The ions in the waters from the Kansas River and upland stream
alluvium with the smallest concentration change with time are bicarbo-
nate and calcium (Fig. 32; 33). In both hydrologic units the bicarbo-
nate concentration changes vary inversely with the calcium concentra-
tion changes suggesting that the bicarbonate and calcium concentration
changes are being dampened by the solubility of calcite.

The concentrations of chloride, calcium, potassium, and sodium of
waters from the upland stream and Kansas River alluvium were lowest
in the sample series taken 3 months after the period of greatest pre-
cipitation (Fig. 32, 33). The concentration changes of these consti-
tuents indicates that the effects of precipitation reach the upland
stream and Kansas River alluvium within three months, as suggested by
Whittemore and others (1982). However, the lowest concentrations of
sulfate in waters from the upland stream and Kansas River alluvium oc-
cur 7 months after the period of the greatest precipitation at a time
when the concentrations of other ions have been increasing. This sug-
gests that it takes up to 7 months for water to infiltrate through the
bedrock and glacial deposits to the alluvium.
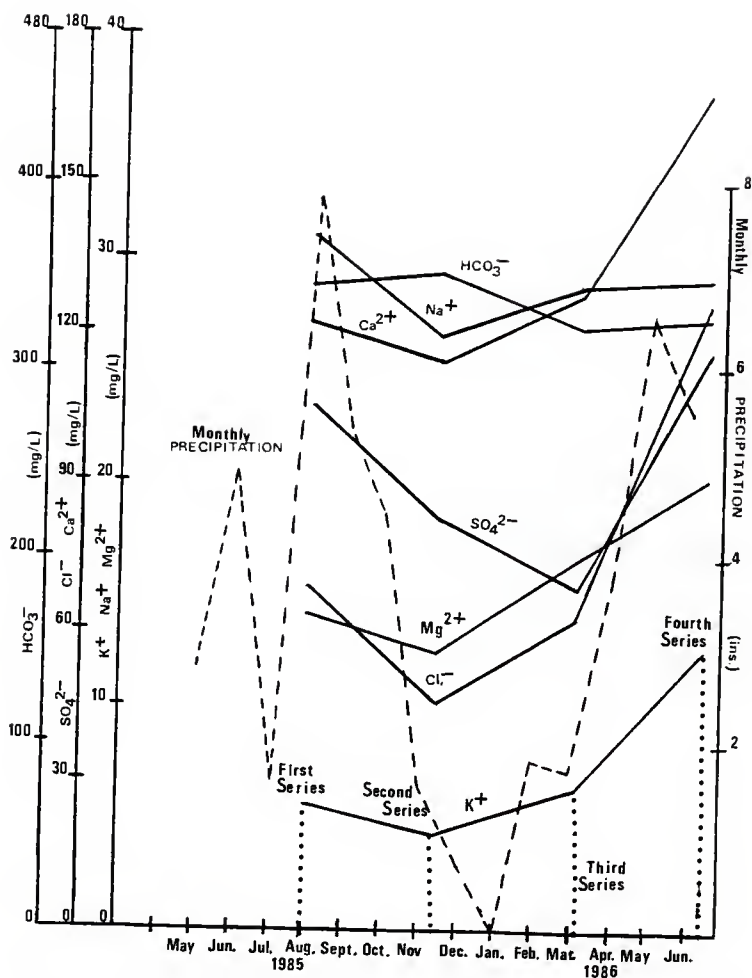
Figure 32. The change in precipitation and in the average composition of water from Kansas River alluvium during the period of investigation.

104
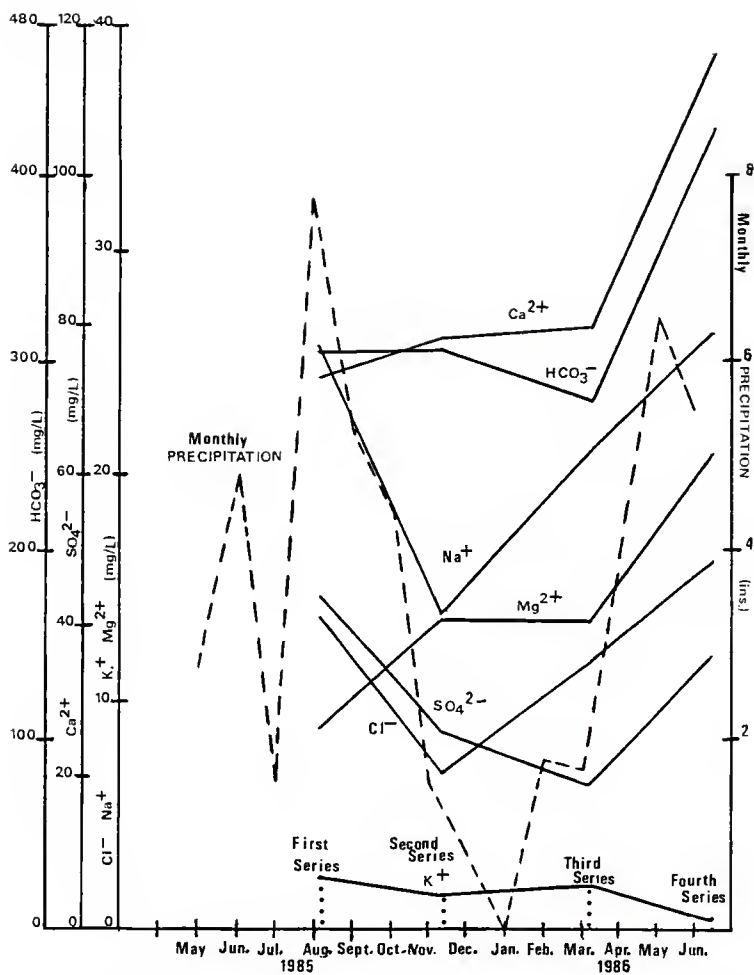
Figure 33. The change in precipitation and in the average composition of water from the upland stream alluvium during the period of investigation.

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<u>Glacial Deposits and Bedrock</u>--The ions in the water from the bedrock
having the greatest relative concentration change with time are sul-
fate, and chloride (Fig. 34). The ions in the water from the glacial
deposits having the greatest relative concentration change with time
are chloride, sodium, and sulfate (Fig. 35). The large variation of
the sulfate concentrations in water from the bedrock and glacial de-
posits appears to be due to the dilution of the sulfate by precipita-
tion. The large chloride concentration change in water from the gla-
cial deposits and bedrock, and the large sodium concentration change
in water from the glacial deposits, might be caused by both variation
in pollution by animal wastes, and by dilution by precipitation.

The ions in the water from the bedrock having the smallest concen-
tration change with time are sodium, bicarbonate, and potassium (Fig.
34). The ions in the water from the glacial deposits having the smal-
lest concentration change with time are calcium, bicarbonate, and po-
tassium (Fig. 35). The small bicarbonate and calcium concentration
changes in the glacial deposits and the small bicarbonate concentra-
tion changes in the bedrock suggest that the precipitation of calcite
may be dampening the calcium and bicarbonate changes. The relatively
constant concentrations of sodium in the bedrock may be due to illite
and montmorillite maintaining a constant concentration of sodium in
the water. The potassium concentrations in the water of the bedrock
and glacial deposits are constant because the bedrock and glacial de-
posits are rarely cultivated.

The magnesium, calcium, sodium, and bicarbonate concentrations in
the water of the bedrock have their lowest concentrations during the
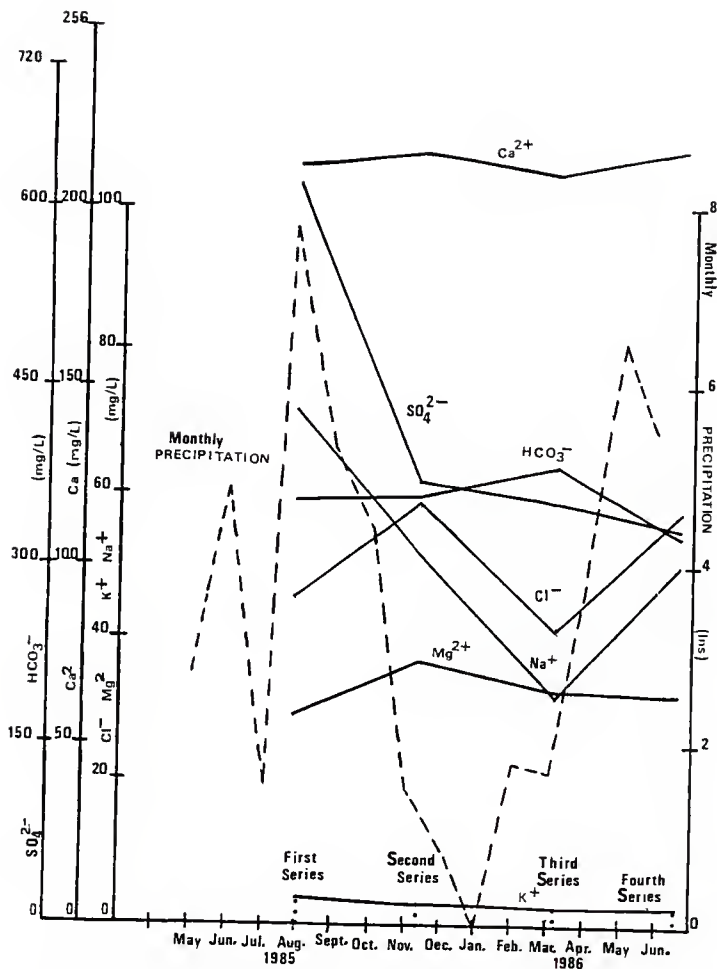
Figure 34. The change in precipitation and in the average composition of water from the glacial deposits during the period of investigation.

107
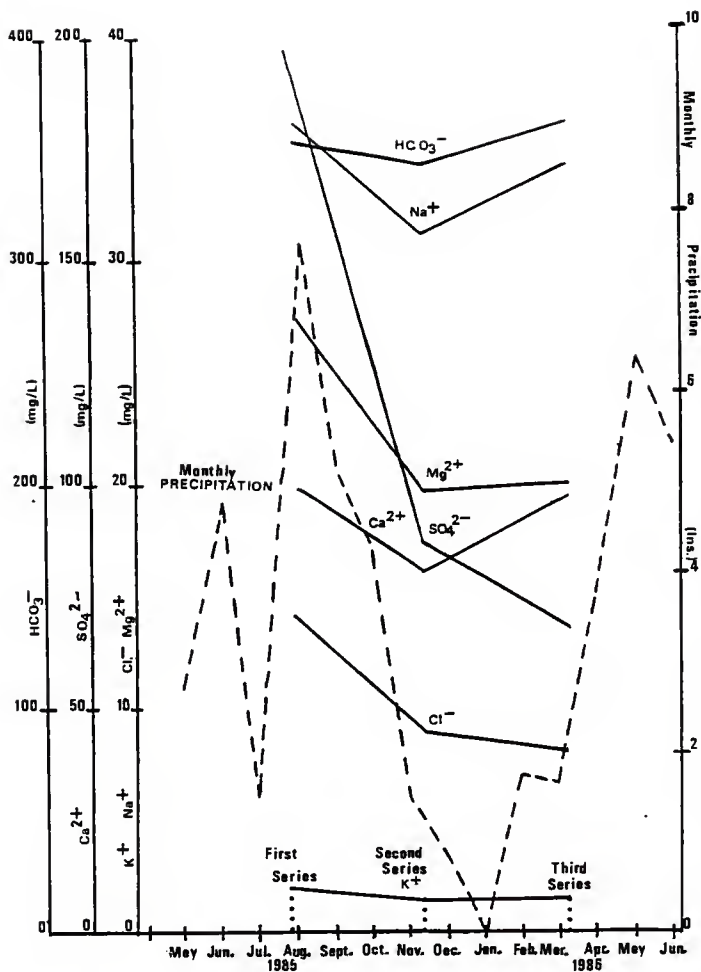
Figure 35. The change in precipitation and in the average compo-
sition of water from the bedrock during the period of
investigation.

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sample series occurring 3 months after the period of highest precipitation (Fig. 34). This suggests that the effects of precipitation can be seen in the quality of water from the bedrock within 3 months and not 6 months to a year, as suggested by Whittemore and others (1982). The difference between the results of this study and the findings of Whittemore and others (1982) are believed to be caused by the shallow depths of wells in the bedrock in the area of investigation.

The bicarbonate, sodium, magnesium, calcium, and potassium concentrations of the glacial deposits have their highest concentrations during the sample series 3 months after the period of greatest precipitation and have their lowest concentrations during the sample series 7 months after the period of greatest precipitation (Fig. 35). The effects of a change of precipitation appear to affect the water quality of the glacial deposits within seven months of a change of precipitation. Whittemore and others (1982) believed that the effects of precipitation were seen in glacial aquifers after long periods of time, and in some cases the effects would not be seen at all. The relatively long period of recharge for the glacial deposits appears to be caused by the small grain size of the deposits causing infiltration of precipitation to be slow.

CONCLUSIONS

The buried, pre-glacial valley in southwest Pottawatomie County is three to five miles wide and is twelve miles long, and extends from one-half mile west of St. George, Ks. to the north-northeast for four and a half miles, trends eastward and runs parallel to Rock Creek until the valleys join half a mile south of Louisville, Kansas. The valleys trend southeast from Louisville and join the Kansas River valley. The gravel and sand deposits of the buried, preglacial valley are a major aquifer in the area and yield between 4 and 450 gallons of water per minute.

The water from the buried, preglacial valley, and most of the water in the area of investigation, is Ca-HCO$_3$ type water containing smaller concentrations of magnesium, potassium, sodium, sulfate, chloride, and trace amounts of strontium, iron, and manganese. The concentrations of all the ions, except strontium and manganese in the water from the pre-glacial, buried valley are significantly lower than the concentrations in the water from other aquifers in the study area. The water from the buried valley had the smallest saturation indices of calcite, gypsum, and dolomite. The low concentration of dissolved solids in the water from the buried valley is due to a lack of significant recharge from the surrounding aquifers and to the buried valley containing insoluble sediments.

Water samples from the glacial deposits contained the highest concentrations of all the ions, except strontium, iron, manganese, and potassium. The water from the glacial deposits had the greatest saturation index of gypsum. The high dissolved solids concentrations of

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water from the glacial deposits are probably due to their receiving mineral-rich recharge from the bedrock. The Kansas River alluvium contains high potassium due to pollution from fertilizers and has the highest average saturation indices of calcite and dolomite. Water from the bedrock contains high strontium concentrations and has low concentrations of potassium, iron, and manganese.

The quality of water from the buried, pre-glacial valley appears to change very little in response to changes in precipitation and the changes occur within three months of the change in precipitation. Changes in precipitation seem to cause changes in the quality of water from the bedrock, Kansas River alluvium, and upland stream alluvium within three months and the changes in the water quality are greater than those in the pre-glacial, buried valley. The quality of water from the glacial deposits appears to change within three to six months of a large change in precipitation.

All the hydrologic units obtain water from the infiltration of precipitation, and the main source of recharge for the buried valley sediments is the infiltration of precipitation. The very low concentrations of sulfate and calcium in the water of the pre-glacial, buried valley suggest that the amount of recharge it receives from other aquifers is small. The aquifers that do recharge the buried valley sediments appear to be the glacial deposits in the western portion of the area and a bedrock, topographic high in the southcentral portion of the area that is bordered on three sides by the buried valley.

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APPENDIX   I

Appendix I. Historical chemical data for groundwaters in the area of investigation.

well location	sample date	temp (C)	spec. cond	pH	Ca2+	Mg2+	Na+	K+	Fe2+	Mn2+	HCO3-	SO42-	Cl-	NO3-	data source
9 10 26db	5/53				130	29	19		1.0		530	17	21	1.0	USGS
9 9 26dd	5/57	14.5	397		47	7.6	30		5.3		240	5.3	5.0	.9	USGS
9 9 27ac	5/57	15.0	273		37	6.7	12		.15		160	.4	4.0	10	USGS
9 9 27cdc	316	8.0	47	5.2	21	1.7	.03	.01	31	---	3.0	1.0			KDHE
9 9 29dc	7/81	16.0	352	5.8	49	4.5	12	2.0				7.0	5.0	3.5	USGS
9 9 32cdd	5/57	14.5	457		58	8.6	29		.92		270	11	5.0	1.0	USGS
9 9 33dbd	7/59		425		58	11	18	12	2.3	.13	250	5.3	12	1.0	USGS
9 9 34aaa	7/86	16.5	175	6.9	20	3.5	11	2.0	.01	---		5.0	5.0	21	USGS
9 10 3cdd	4/53				62	10	24		.28		140	64	34	18	USGS
9 10 9bca1	5/85		662	7.8	106	12	35	.8	1.9	.33	330	29	14	.02	KDHE
9 10 9bca2	9/79		730	7.6	114	14	40	1.0	1.5	.17	382	27	11	.4	KDHE
9 10 9cab	1/86		287	7.7	39	5.0	13	1.8	.22	.02		25	6	8.5	KDHE
9 10 26db	4/53				130	29	19		1.0		533	17	21	1.0	Beck, 1959
10 8 1cb	5/57	14.5	501		65	11	25				300	.4	7	2.0	USGS
10 8 12cbb	9/66	15.0	870	7.8	140	15	27	10	12.	.38	470	71	18	1.8	USGS
10 8 13bdc1	3/67	18.0	640	7.7	120	12	10	5.6	3.3	2.0	390	26	12	.9	USGS
10 8 13bdc2	8/67	18.0	690	7.5	120	13	12	5.2	2.9	2.0	410	32	13	1.3	USGS
10 8 14cba	3/67	18.5	900	7.5	110	19	57	6.0	4.3	.86	290	89	100	1.5	USGS
10 9 1ccd1	5/57	14.5	532		66	8.6	28		.53	---	220	50	15	5.8	USGS
10 9 1ccd2	7/59	14.5	725		92	15	43		1.0	1.0	330	54	35	5.3	USGS
10 9 7bbd	12/85		190	7.7	22	2.9	12	1.2	---	---		---	2.5	2.2	KDHE
10 9 9cad	10/85		290	7.2	37	5.7	15	2.8	.30	.05		19	8.3		KDHE
10 9 9cbc	12/85		350	7.6	56	6.0	13	1.6	---	---		18	7.9		KDHE
10 9 9cda1		15.0	400	6.9	56	6.0	16	2.0	.24	---	170	16	13	31	USGS
10 9 9cda2	4/63		360	7.3	55	4.6	13	1.7	.63	.12	180	8.6	9.0	4.1	
10 9 9cda3	2/74		370	7.1	54	6.2	14	1.8	---	---	190	15	10	3.6	
10 9 9cdc1	9/79	18.5	360	6.3	42	5.6	17	2.0	---	---	150	16	10	5.3	USGS
10 9 9cdc2	6/85	14.5	324	5.6	58	5.9	13	1.8	.01	.01		13	7.4	---	
10 9 14dcb1	5/57	14.5	815		140	14	17	---	23	---	470	38	10	1.6	USGS
10 9 14dcb2	3/67	15.0	530	7.5	100	12	11	4.4	---	.22	330	43	11	.4	
10 10 1bca	6/66	15.0	590	7.4	93	12	17	10	2.4	.19	280	24	9.0	1.3	USGS
10 10 2bcc	8/67	14.5	560	7.3	74	10	34	8.0	3.7	.27	320	20	17	2.2	USGS
10 10 3ccb1	10/67		980	7.0	130	19	51	9.3	.60		250	170	99	1.3	USGS
10 10 3ccb2	3/68		930	6.6	120	22	47	10	.28		220	170	88	7.1	
10 10 3cd	4/53		307		62	10		24	.28			64	34	18	Beck, 1959
10 10 5ddc	7/85				97	12	23	1.6	.03	.01	222	67	35	9.3	KDHE
10 10 9aca1	3/62		830	7.3	91	18	60	---	.79	.45	270	86	82	2.2	USGS
10 10 9aca2	4/64		830	7.3	96	16	70	7.3	.49	.57	280	89	95	1.5	
10 10 9aca3	10/66		750	8.0	78	15	52	7.8	.27	.62	240	74	75	3.5	
10 10 9bdb1	7/83	14.5	834	7.5	91	15	60	6.0	.03	.77	250	110	68	2.7	USGS

well location	sample date	temp (C)	spec. cond	pH	Ca2+	Mg2+	Na+	K+	Fe2+	Mn2+	HCO3-	SO42-	Cl-	NO3-	data source
10 10 9bdb2	5/84	15.5	783	7.1	110	17	53	6.5	.02	.99		110	66	2.3	
10 10 9bdb3	6/85	17.0	822	6.9	110	18	49	6.1	.04	.72		110	64	3.0	
10 10 10bdc	3/67	14.0	630	7.8	72	13	27	10	.02	.11	230	40	30	.9	USGS
10 10 10dbc	6/66	14.0	670	7.5	82	16	32	6.6	1.7	2.7	280	59	46	1.8	USGS
10 10 11aac	6/66	14.5	750	8.1	110	4.7	35	8.0	.61	.31	320	68	49	.9	USGS
10 10 11dc	5/53		488		130	16		22	.29	---	370	55	24	28	Beck, 1959
10 10 12bcd	8/67	15.0	770	7.5	120	10	2.9	7.8	2.8	.48	360	66	41	1.3	USGS
10 10 16ac	5/51		413		109	14		25	.14		394	38	10	8.8	Beck, 1959
10 10 18acc	8/68	14.0	610	7.4	96	16	16	5.4	6.0	.68	340	48	12	2.7	USGS
10 10 19bac1	5/66	14.5	800		90	10		7.0	3.2	.86	260	75	82	1.8	USGS
10 10 19bac2	3/67	15.0	670		74	7.7		5.9	2.3	.64	260	58	53	.4	USGS
10 10 19cca		14.5	700	7.4	98	7.7	42	5.4	1.2	.75	280	66	59	4.2	USGS

Note: The concentrations are expressed in milligrams per liter.
Note: the specific conductivity is expressed in micromhos per centimeter at 25 C.

APPENDIX   II

Appendix II. Well sites visited during the investiga-
tion.

well location	well owner	well location	well owner
8 9 35cdd	O. Blanka	9 10 35dca	R. Peddicord
		9 10 35dbb	R. Peddicord
9 8 2ddd	D. Kunidger	9 10 32dbb	C. Chapman
9 8 25dcd	D. Irwin	9 11 6bcc	R. Peterson
9 8 26dbd	E. Thiele	9 11 6cba	R. Peterson
9 8 11bdd	C. Bellinger		
9 8 26aba	M. Briggs	10 8 2bcd	T. Ubel
9 8 26bcc	M. Dikeman	10 8 11dad	C.K. Dehydrato
		10 8 12cab	K. Kern
9 9 4ddd	F. Ebert	10 8 12dba	C. Cragg
9 9 4ddc	F. Ebert		
9 9 7ddd	B. Heptig	10 9 2add	W. Hildreth
9 9 11daa	Bartholomew	10 9 3dad	M. Anderson
9 9 15dbc	W. Patnode	10 9 3ddc	D. Altwegg
9 9 24add	F. Warnow	10 9 4cab	M. Crubel
9 9 28dcd	D. Danker	10 9 4dcd	E. Lundquist
9 9 29cdb	PCRWD #1	10 9 5bbb	C. Meyer
9 9 29cdc	PCRWD #1	10 9 5ddc	L. Carver
9 9 29dcc	PCRWD #1	10 9 6ada	C. Meyer
9 9 30aaa	G. Burnett	10 9 6bbb	S. Torrey
9 9 30aad	D. Zarger	10 9 7acc	J. Wilson
9 9 30abd	D. Westgate	10 9 7bad	D. Torrey
9 9 31baa	D. Ubel	10 9 8aca	J. Wilson
9 9 33	C. Meyer	10 9 8bbd	V. Claeys
9 9 33ccc	K. Piper	10 9 8dac	G. Heideman
9 9 34aab	J. Lintz	10 9 9cad	St. George
9 9 35aaa	T. Cunningham	10 9 9bdc	D. Burgess
9 9 36bcc	L. Hudson	10 9 10baa	T. Vera
		10 9 10caa	G. Smith
9 10 3cdd	L. Blanka	10 9 10bbc	A. Virgin
9 10 6aab	D. Brunkow	10 9 10ddd	M. Eschwein
9 10 9bbb	Pott Co.	10 9 11cca	M. Toy
9 10 16cac	D. Umschied	10 9 11ccc	J. Estes
9 10 20baa	L. Brayton	10 9 12abb	M. Toy
9 10 21ccc	T. Picolett	10 9 14adc	H. Johnson
9 10 21dbb	V. Robinson	10 9 14dac	R. Shaw
9 10 28bda	L. Bammes	10 9 14dbc	W. Taylor
9 10 29dcc	Pott. Count.	10 9 14dbc	H. Johnson
9 10 30dca	L. Peddicord	10 9 14dbb	L. Johnson
9 10 31aaa	J. Winterman	10 9 15abd	D. Drouillard
9 10 32bba	O. Richards	10 9 15aca	P. Gardner
9 10 32bcc	Snapp-Wilson	10 9 15bbb	M. Oppenlander
9 10 32cbc	A. Seele		
9 10 32dbd	J. Waters	10 10 1bda	L. Sylvestor
9 10 33bbd	I. Duncan	10 10 2cbc	Dr. L. Otto
9 10 35bbc	E. Boyce	10 10 2dcb	D. Crest

well location	well owner	well location	well owner
10 10 3bab	L. Weixelman	10 10 10aba	R. Fairchild
10 10 3ddd	Q. Carnahan	10 10 10abb	R. Fairchild
10 10 5bbb	K. Lamb	10 10 10bbb	Wamego,
10 10 6baa	D. Douglas	10 10 11baa	D. Carnahan
10 10 6bbb	D. Butler	10 10 17bbb	M. Grothjan
10 10 6bbd	Huppe	10 10 17caa	D. Doperalski
10 10 6bdd	J. Hildreth	10 10 17ddc	D. Doperalski

APPENDIX   III

Appendix III. Results of the chemical analyses.

well location	well owner	sample date	temp. (C)	spec. cond	pH	Ca2+	Mg2+	Na+	K+	Sr2+	Fe2+	Mn2+	HCO3-	SO42-	Cl-	NO3-
8 9 35 cdd#1	O. Blanka	7/29/85	19.6	1465	7.0	110	35	85	3.7	26	DL	DL	447	519	16	
8 9 35 cdd#2	O. Blanka	11/12/85	13.9		6.9	113	34	81	2.5	26	.30	DL	457	171	10	13
8 9 35 cdd#3	O. Blanka	3/6/86	11.3	998	7.0	93	33	74	2.7	26	DL	DL	403	222	13	8
9 8 26dbd#1	Thiele	8/1/25	14.5	1240	7.1	118	12	14	1.1	DL	DL	DL	332	56	21	
9 8 26dbd#2	Thiele	11/13/85	15.8	865	6.9	121	18	12	.80	1.8	.10	DL	398	30	10	46
9 8 26dbd#3	Thiele	3/4/86	14.9	724	7.0	87	17	11	1.0	1.0	DL	DL	355	73	12	54
9 9 4ddd#1	Ebert	8/8/85	17.6	986	7.2	130	15	27	2.4	1.3	4.4	1.3	427	77	14	
9 9 4ddd#2	Ebert	11/8/85	14.2	9267	7.1	114	23	25	1.6	1.4	4.1	.17	450	49	13	12
9 9 4ddd#3	Ebert	3/7/86	11.8	6777	7.2	119	22	24	2.1	1.0	3.1	1.2	393	35	13	6
9 9 4ddd#4	Ebert	6/17/86	16.8	811	6.8	116	21	26	2.2	1.1	13.8	DL	427	36	16	8
9 9 11daa	Bartholomew	7/27/85	20.1	1024	6.8	129	11	11	.80	.80	DL	DL	447	DL	7.0	
9 9 15dbc#1	Patnode	7/26/85	19.5	646	7.4	74	11	21	3.3	.36	.63	.75	286	27	5.0	
9 9 15dbc#2	Patnode	11/13/85	14.5	593	7.3	70	12	20	2.4	.37	.50	.12	272	22	4.0	DL
9 9 15dbc#3	Patnode	3/6/86	7.4	401	7.1	72	11	23	2.3	.25	.40	.70	292	19	2.0	DL
9 9 24add#1	Warnow	7/29/85	15.9	346	7.1	35	11	17	2.2	.14	DL	DL	162	10	8.0	
9 9 24add#2	Warnow	11/12/85	13.8	375	6.7	35	3.2	17	1.6	.26	.2	DL	160	1.0	4.0	7
9 9 24add#3	Warnow	3/6/86	13.7	241	6.8	39	3.4	15	1.4	.10	DL	DL	172	2.0	2.0	3
9 9 24add#4	Warnow	6/18/86	16.0	298	7.1	37	3.2	17	1.8	.11	DL	DL	188	2.0	3.5	DL
9 9 31baa#1	D. Ubel	3/6/85	16.3	629	7.3	65	9.1	37	3.3	.57	DL	DL	285	9.0	4.0	
9 9 31baa#2	D. Ubel	11/7/85	14.6	459	7.3	42	6.7	33	2.4	.38	.30	.06	256	3.0	2.0	10
9 9 31baa#3	D. Ubel	3/7/86	7.3	357	6.9	50	7.8	16	2.4	.35	DL	DL	250	2.0	1.0	6
9 9 31baa#4	D. Ubel	6/15/86	14.7	495	7.2	54	8.5	33	2.7	DL	DL	DL	292	DL	1.6	10
9 9 33adb#1	Piper	7/26/85	15.6	134	6.9	21	2.5	17	2.7	.13	DL	DL	93	8.0	4.0	
9 9 33adb#2	Piper	11/12/85	13.9	214	6.8	14	2.7	14	2.0	.13	.10	DL	89	DL	1.0	12
9 9 33adb#3	Piper	3/4/86	13.8	86	6.7	15	2.0	9.0	1.1	.10	DL	DL	101	2.0	1.0	5
9 9 33adb#4	Piper	6/15/86	14.9	208	7.1	21	2.8	14	2.2	.12	DL	DL	118	DL	4.4	9
9 9 34abb	Lintz	7/27/85	19.3	197	6.8	23	2.4	13	2.3	.12	DL	DL	77	8.0	5.0	
9 9 36bcc	Hudson	7/30/85	15.4	357	6.9	24	2.7	20	1.8	.17	DL	DL	130	9.0	5.0	
9 10 3ddd	L. Blanka	7/29/85	18.9	3948	6.8	324	32	140	1.1	1.7	.32	.10	353	44	16	
9 10 6aab	Brunkow	7/29/85	13.9	1478	7.2	130	31	41	2.4	3.5	DL	.20	188	DL	38	
9 10 14aba	Whearty	8/5/85	19.5	1537	7.3	118	27	63	3.3	3.8	5.1	2.4	511	86	67	
9 10 16cac#1	Umschied	7/30/85	17.9	1657	6.8	119	28	117	1.6	DL	DL	.16	276	272	73	
9 10 16cac#2	Umschied	11/11/85	14.0	1926	6.2	143	32	89	2.1	.59	.40	DL	286	181	103	192

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well location	well owner	sample date	temp. (C)	spec. cond	pH	Ca2+	Mg2+	Na+	K+	Sr2+	Fe2+	Mn2+	HCO3-	SO42-	Cl-	NO3-
9 10 16cac#3	Umschied	3/6/86	14.0	973	6.7	117	25	44	1.6	.29	.13	.14	239	92	69	180
9 10 16cac#4	Umschied	6/18/86	13.8	1249	7.1	137	33	75	1.5	.55	.13	.10	286	115	95	
9 10 28bda#1	Bammes	7/24/85	15.1	345	7.0	56	2.6	24	2.0	.18	DL	DL	179	11	13	
9 10 28bda#2	Bammes	11/8/85	15.0	365	6.6	41	4.3	1.6	1.7	.14	.2	DL	168	3.0	10	13
9 10 28bda#3	Bammes	3/5/86	10.7	292	6.5	40	4.9	19	1.5	.15	DL	DL	173	7.0	10	7
9 10 32cbc#1	Seele	7/29/85	16.6	552	7.6	46	5.0	35	2.1	.22	DL	DL	245	8.0	10	
9 10 32cbc#2	Seele	11/8/85	4.3	477	7.1	50	4.7	33	1.8	.17	.16	.06	258	1.0	6.0	11
9 10 32cbc#3	Seele	3/5/86	14.0	430	7.1	49	4.8	31	1.7	.19	DL	DL	224	2.0	12	5
9 10 35bbc	Boyce	8/2/85	14.6	918	7.1	99	13	45	5.1	.5	2.8	.19	381	26	15	
9 11 6bcc	Peterson	8/8/85	18.4	1742	7.1	177	32	64	3.7	1.3	DL	.54	455	247	65	
10 8 2bcd	T. Ubel	8/6/85	15.5	419	7.3	72	6.4	19	2.3	.41	DL	DL	263	12	9	
10 8 12cab#1	Kern	8/1/85	13.8	671	7.1	86	10	27	4.4	.37	DL	DL	344	20	10	
10 8 12cab#2	Kern	11/13/85	13.1	710	7.0	91	9.3	25	3.2	.40	4.1	DL	374	9.5	6	2
10 8 12cab#3	Kern	3/4/86	13.8	596	7.1	81	9.1	25	3.2	.26	DL	DL	321	14	6	1
10 9 3ddc#1	Altwegg	8/6/85	15.0	488	7.5	68	7.0	9.6	1.4	.23	.24	.10	232	17	6	
10 9 3ddc#2	Altwegg	11/13/85	14.3	419	7.1	61	6.6	9.5	1.1	.19	.10	DL	228	7	3	10
10 9 3ddc#3	Altwegg	3/10/86	14.0	386	7.2	61	8.0	6.5	.8	.16	.14	DL	223	14	2	8
10 9 5ddc#1	Carver	8/1/85	14.8	397	6.8	34	3.9	20	2.5	.19	DL	DL	119	19	2	
10 9 5ddc#2	Carver	11/8/85	14.6	426	7.0	32	.82	21	.7	DL	.40	DL	124	1	10	48
10 9 5ddc#3	Carver	3/10/86	14.3	323	6.8	49	6.6	13	2.9	.21	DL	DL	108	14	13	62
10 9 5ddc#4	Carver	6/17/86	14.3	347	6.8	49	5.4	12	2.4	.22	DL	DL	112	13	14	57
10 9 6ada	Meyer	8/1/85	15.6	320	7.2	30	3.4	14	2.3	.16	DL	DL	93	19	5	
10 9 7baa	Torrey	7/23/85	15.1	129	6.5	10	.70	17	1.5	DL	DL	DL	56	17	5	
10 9 14dac	Shaw	7/30/85	18.7	977	7.2	113	5.8	12	3.7	.31	DL	DL	291	19	13	
10 9 15bbb#1	Oppenlander	7/23/85	16.8	1957	7.0	306	29	25	4.4	.51	DL	DL	428	970	14	
10 9 15bbb#2	Oppenlander	11/8/85	14.1	2286	6.4	287	39	14	3.0	.42	.20	DL	427	559	13	12
10 9 15bbb#3	Oppenlander	3/7/96	11.0	1496	6.8	302	39	20	2.9	.33	DL	DL	400	512	12	8
10 9 15bbb#4	Oppenlander	6/16/86	14.8	1348	6.7	295	37	25	3.1	.37	DL	DL	432	549	18	7
10 10 17cac	Doperalski	7/27/85	15.6	868	7.4	120	5.7	18	5.3	.39	2.9	.70	325	102	16	
10 10 1bda	Sylvestor	7/27/85	14.8	972	7.2	100	7.2	19	15	.50	4.5	.20	370	19	18	
10 10 6bdd#1	Hildreth	7/24/85	15.9	375	6.7	28	4.9	25	1.4	DL	DL	DL	136	92	10	
10 10 6bdd#2	Hildreth	11/11/85	14.9	391	6.8	30	4.3	22	1.2	.29	.30	DL	124	8	7	11
10 10 6bdd#3	Hildreth	3/4/86	13.7	253	6.4	25	4.0	21	1.0	.10	DL	DL	105	11	8	20

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well location	well owner	sample date	temp. (C)	spec. ccnd	pH	Ca2+	Mg2+	Na+	K+	Sr2+	Fe2+	Mn2-	HCO3-	SO4=-	Cl-	NO3-
10 10 10abb#1	Fairchild	7/24/85	14.7	1434	7.3	170	19	35	26	.70	DL	DL	350	169	45	
10 10 10abb#2	Fairchild	11/7/85	14.6	1041	7.2	137	18	28	17	.60	.20	.35	329	158	33	62
10 10 10abb#3	Fairchild	3/5/86	14.0	1250	7.0	175	24	31	25	.57	DL	DL	324	121	51	220
10 10 10abb#4	Fairchild	6/15/86	14.6	893	7.2	169	20	29	28	.58	.10	DL	328	116	56	

Note: The concentrations are expressed in milligrams per liter.
Note: the specific conductivity is expressed in micromhos per centimeter at 25 C.

APPENDIX   IV

Appendix IV.   Procedures for Chemical Analysis.

General

Solutions of low concentrations were prepared just be-
fore they were to be used to avoid absorption on the side
of the containers. The same volumetric flask was used to
prepare all the samples for a particular analysis. Automa-
tic pipettes were used to prepare the standards for the
analyses of the cations. The automatic pipette tips were
placed on the pipettes without being touched by a hand. The
pippette tips were not placed on the counter or anyplace
where they could become contanimated. The pipette tips were
disposed of after the analyses in which they were used.

All beakers, flasks, pipettes, and sample bottles used
in the analyses were rinsed by tap water and washed with
soapy water and a wire brush. The containers were rinsed
several times with tap water, and then rinsed several times
with distilled water, and given a final series of rinses
with D-D water (distilled and deionized water). Sample bot-
tles, pipettes and cuvettes were rinsed several times with
the sample water before they were used for the analysis.
When a magnetic stirrer was used in an analysis, the stir-
ring speed was kept constant for that series of analyses.
The beakers and the flasks were all dried before they were
used in an analysis.

Reagent or primary standard chemicals were used for all
the analyses. the solid chemicals were dried in a drying
oven for several hours at 200 C, and allowed to cool to
room temperature within a desicator.

Bicarbonate (HCO_3^-)

Method: Potentiometric Titration (American Public Health

Association, 1975)

Apparatus: 50 ml buret, Corning model 610 pH/mV-meter,

Magnetic stirrer, pH-electrode.

Reagents: Concentrated H_2SO_4 (sulfuric acid),

Na_2CO_3 (sodium carbonate).

Procedure:

1. Calibrate the pH-meter to pH 4.0 and pH 7.0 according
    to the instructions found in the pH-meter manual
    (Corning Scientific Instruments, 1972).

2. Prepare a 0.1 N (normal) sulfuric acid solution by
    diluting 3.0 ml of concentrated sulfuric acid to 1.0
    liter with D-D water.

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3. Prepare a 0.02 N sulfuric acid solution by diluting 200 ml of 0.1 N sulfuric acid solution to a volume of 1.0 liter. The 0.02 N sulfuric acid is to be used for titrating samples with 20 mg/l or less of bicarbonate.

4. Prepare a 0.05 N sodium carbonate standard by weighing out 2.500 g (grams) of sodium carbonate, dissolving it and diluting it to 1.0 liter using D-D water. The sodium carbonate standard must be prepared the day when it will be used.

5. Fill the 50 ml buret with the 0.10 N sulfuric acid and place the buret where it may be used to titrate the sodium carbonate standard.

6. Pipette 25 ml of .05 N sodium carbonate standard into a plastic beaker containing a magnetic stirring rod. Place the beaker on the magnetic stirrer and turn the magnetic stirrer on.

7. Rinse the pH-electrode with D-D water and wipe it dry with a chem-wipe. Place the pH-electrode into the support arm of the pH-meter and immerse the glass bulb of the pH-electrode in the sodium carbonate.

8. Make sure the magnetic stirring bar doesn't strike the glass bulb of the pH-electrode. Turn on the magnetic stirrer and the pH-meter.

9. Titrate the sodium carboante with the sulfuric acid until the sodium carbonate solution is at pH 4.5. Record the level of the sulfuric acid in the buret.

10. Repeat steps five through nine substituting the sample waters for the sodium carbonate.

Calculations:

1. Calculate the normality of the sulfuric acid where;

$$N = \frac{A\ (B)}{53.00\ (C)}$$

Where,
N = normality of the sulfuric acid
A = the grams of sodium carbonate
B = ml of the sodium carbonate solution used for the titration.
C = ml of sulfuric acid solution used for the titration.

2. Calculate the bicarbonate (mg/l) using the equation;

$$A = \frac{(B)\,60.02\ (C)\,1000}{S}$$

Where, A = bicarbonate of the sample

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B = normality of the sulfuric acid used to
    titrate the sample
C = amount of sulfuric acid used to
    titrate the sample
S = milliliters sample used

Standard deviation: 4.0 Percent (American Public Health
                    Association, 1975).

Chloride (Cl^-)

Method: Argentometric   Titration   (American   Public   Health

        Association, 1975).

Apparatus: 50 ml buret, magnetic stirrer, funnel, filter

           paper, eye-dropper.

Reagents: K_2CrO_4 (Potassium chromate), $AgNO_3$ (silver ni-

          trate), NaCl (sodium chloride).

Procedure:
1. Prepare a silver solution by dissolving g silver
   nitrate in liter water. Store the silver solution in a
   brown bottle.

2. Prepare a potassium chromate indicator solution by dis-
   solving 50 g of potassium chromate into 1.0 liter of D-D
   water.

3. Add silver nitrate to the potassium chromate indicator
   solution until a red precipitate forms. Let the potas-
   sium chromate indicator solution stand for 12 hours and
   filter the potassium chromate solution.

4. Prepare a .0141 N sodium chloride solution by dissolving
   .8214 g of sodium chloride into 1.0 liter of D-D water.

5. Fill the 50 ml buret with silver nitrate and position
   the buret over the beaker with sodium chloride standard.

6. Pipette 25 ml of .0141 N sodium chloride solution into a
   glass beaker containing a magnetic stirring bar.

7. Add five drops of potassium chromate indicator solution
   to the sodium chloride in the beaker and place the
   beaker on the magnetic stirrer. Turn on the magnetic
   stirrer.

8. Titrate the sodium chloride standard with the silver
   nitrate solution until the sodium chloride standard
   becomes pinkish-yellow, and the color remains for more

than 15 seconds. Do not throw the sodium chloride stan-
dard away.

9. Record the amount of silver nitrate solution used in the
titration and make a note of it.

10. Repeat steps five through nine using a blank of 25 ml
of D-D water. Try to match the final color of the blank
with the final color of the sodium chloride standard.

11. Repeat steps five through nine using 25 ml of sample
water. Try to match the color of the sample to the final
color of the sodium chloride standard.

12. The red precipitate will gradually settle out of the
sodium chloried standard. Run a new sodium chloride
standard and a new D-D water blank every six samples.

Calculations:

1.Calculate the normality of the silver nitrate solution by
using the equation;

$$N = \frac{.0141 \ (25 \ ml)}{B - C}$$

Where;   N = normality of the silver nitrate solution
         B = ml of silver nitrate used to titrate the
                 sodium chloride standard
         C = ml of silver nitrate solution used to titrate
                 the blank

2.   Calculate the normality of the chloride in the sample
water by using the equation;

$$S = \frac{35.45N \ (B - C) \ 1000}{25 \ ml}$$

Where, S = normality of the sample
        N = normality of the silver nitrate solution
        B = Amount of silver nitrate solution used to
            titrate the sample water
        C = Amount of silver nitrate solution used to
            titrate the blank.

Standard Deviation: 4.2 percent (American Public   Health
                        Association, 1975)

Sulfate (SO_4^{2-})

Method: Turbidimetry (American Public Health Association,

1975)

Apparatus: Magnetic stirrer, Coleman Universal Spectro-
photometer Model-14, Spectrophotometer Cuvette, stop watch.

Reagents: Na_2SO_4 (sodium sulfate), 30-40 mesh $BaCl_2$ (barium
chloride) crystals, glycerol, isopropyl alcohol,
NaCl (sodium chloride), concentrated HCl.

Procedure:

1. Prepare a 100 mg/l sulfate solution by dissolving .1479 g of sodium sulfate into 1.0 liter of D-D water.

2. Prepare 5, 10, 15, 20, 30, and 40 mg/l sulfate standards from the 100 mg/l sulfate solution by diluting 5, 10, 15, 20, 30, and 40 ml. of the 100 mg/l sulfate solution to 100 ml.

3. Prepare three sets of standards by pouring 25 ml of each standard into three different beakers. Seal the beakers tightly with cellophane and a rubber band.

4. Prepare the samples for the analysis by pipetting 25 ml of each sample into a beaker and sealing it with cellophane and a rubber band. Prepare three beakers for each sample.

5. Prepare a blank for setting the spectrophotometer and one blank for every five samples and standards that have been prepared for the analyses by pipetting 25 ml of D-D water into a beaker and sealing it with cellophane and a rubber band.

6. Prepare the spectrophotometer to be used for turbidimetry by following the instructions in the operating manual (Coleman Instruments, 1968). Set the spectrophotometer at 420 nanometers and allow the spectrophotometer to warm up for at least an hour. The samples and standards must be at the same temperature for the analysis.

7. Prepare a salt-acid glycerol conditioning reagent by dissolving 75 g of sodium chloride in a solution consisting of 300 ml of D-D water, 30 ml of concentrated hydrochloric acid and 100 ml of isopropyl alcohol. Mix 50 ml of glycerol into the acid-salt solution after all the sodium chloride has been dissolved.

8. Place a magnetic stirring bar into a beaker containing a blank, and place the beaker on the magnetic stirrer and turn on the magnetic stirrer. The magnetic stirrer must be at the same stirring speed for the set of analyses and the magnetic stirring bars should be the

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same size and shape.

9. Add 2.0 ml of conditioning reagent to the blank, and add a small amount of barium chloride crystals to the beaker. The same amount of barium chloride crystals should be added to each sample and standard.

10. Start the stop watch as the barium chloride crystals are being added to the blank and stir the solution for exactly one minute.

11. Immediately rinse the spectrophotometer cuvette three times with the blank solution, wipe the outside of the cuvette clean with a chem-wipe, and place the cuvette in the spectrophotometer light beam with the trade mark of the cuvette facing the top of the spectrophotometer.

12. Set the galvanometer scale to 100 percent transmittance, and allow the cuvette to remain in the spectrophotometer for six minutes after the barium chloride crystals were added to the blank.

13. Check the galvanometer scale to ensure that it is at 100 percent transmittance. This setting is the blank value to compare the sample and standards against.

14. Take the cuvette out of the spectrophotometer and rinse it. Carefully set the cuvette to one side.

15. Repeat steps 8 through 11 using either a sample or a standard. A blank should be run every five samples and the galvanometer reset.

16. Once the cuvette is placed in the spectrophotometer light beam the percent transmittance should be recorded for the samples and the standards at half minute intervals until six minutes have passed since the BaCl was added to the solution.

17. If a sample has a reading which has a lower transmittance than the 40 mg/L sulfate standard, then the sample should be diluted to a concentration with a higher transmittance than that of the 40 mg/l sulfate.

18. Clean the cuvette throughly after each analysis.

Calculations:

1. The values for the standards of the same concentration should be added together and an average used for the calculations.

2. Use semi-log paper to prepare a standard curve. Plot the sulfate concentration on the arithmetic scale and plot the percent transmittance on the logarithmic scale (Fig. 36). Draw a straight line through the points.

a.

Sulfate Standard Concentration (mg/l)	Percent Transmittance	Sulfate Standard Concentration (mg/l)	Percent Transmittance
0	100	20	69
5	91	30	55.5
10	83.5	40	46.5
15	75	Sample	52

b.
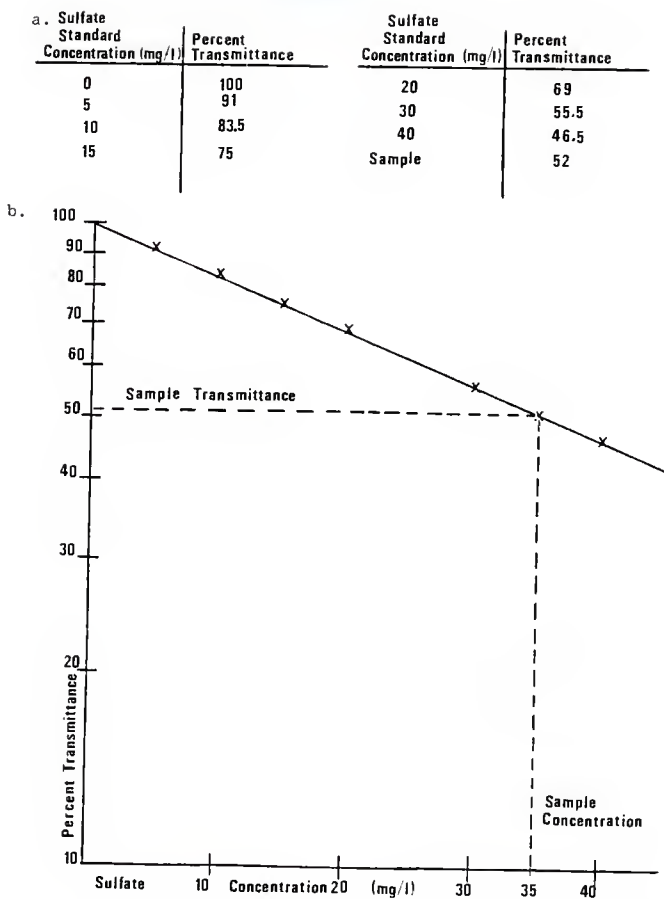
c. Sulfate concentration = sample concentration (dilution)

Fig. 36 . The steps used in calculating the sulfate concentrations:
a.   Table of the sulfate standards and samples transmittance. b. The plot of the data in part a. c. Calculations of the actual sulfate concentrations.

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3. Read the sulfate concentrations from the curve and multiply the sulfate concentration by the samples dilution factor to obtain the samples sulfate concentration.

Standard Deviation: 9.1 Percent (American Public Health Association, 1975).

Nitrate (NO_3^-)

Method: Ion specific electrode (Orion Research Incorporated, 1974)

Equipment: Orion Nitrate ion specific electrode, Beckman calomel reference electrode, Corning model 610 pH/mV meter, magnetic stirrer,

Reagents: $NaNO_3$ (sodium nitrate), H_2SO_4 (sulfuric acid), $(NH_4)_2SO_4$ (ammonium sulfate).

Procedure:

1. Prepare the nitrate ion electrode according to the instructions in the manufacturers manual (Orion Research Incorporated, 1974). Prepare the pH/mV meter to be used for millivolts according to the manufacturers manual (Corning Scientific Instruments, 1972).

2. Prepare a 2.0 N ionic strength adjusting solution by dissolving 26.4 g of ammonium sulfate into 100 ml of D-D water.

3. Prepare a .4 N ionic strength adjusting solution by diluting 20 ml of 2.0 N ionic strength adjustor to 100 ml using D-D water.

4. Prepare a 10,000 mg/l nitrate solution by dissolving 13.14 g of sodium nitrate into 1.0 liter of D-D water.

5. Prepare a 1,000 mg/l nitrate solution by diluting 100 ml of 10,000 mg/l nitrate solution to 1.0 liter using D-D water.

6. Prepare 100 and 10 mg/l nitrate solutions by diluting 100 ml of 1,000 mg/l nitrate solution to 1.0 liter and by diluting 100 ml of 100 mg/l nitrate solution to 1.0 liter.

7. Preserve the standards with the same treatment that was used to preserve the samples. The nitrate standards must be made the same day that the analysis is run.

8. Measure out 50 ml of standards and samples into plastic beakers and add 2.0 ml of ionic strength adjustor to each of your samples and standards.

9. Add the same amount of sulfuric acid to each sample and standard. The pH of all the samples must be below 4.5.

10. Place a clean magnetic stirring bar into a beaker, and place the beaker on a magnetic stirrer.

11. Turn on the magnetic stirrer and mix the solution well. Turn off the magnetic stirrer.

12. Rinse the electrode assembly with D-D water, dry it carefully with a chem wipe and place the electrode assembly into the beaker. Make sure no air bubbles are adhering to the electrode assembly.

13. Turn on the pH/mV-meter and take periodic readings of the millivolts until the readings are constant.

14. Turn off the pH/mV-meter and lift the electrode out of the beaker. Repeat steps 10 through 15 with a new beaker. Repeat each beaker three times.

15. Let the nitrate specfic ion electrode stand in 1,000 mg/l nitrate standard during the time it will be needed for use. Once the electrode is no longer needed it should be cleaned and stored.

Calculations:

1. Average the two closest millivolt readings of each sample and standard to attain the reading to be used for calculations.

2. Plot the standards on semi-log graph paper with the concentration on the logarithmic axis and the millivolts on the arithmetic axis (Fig. 37). Draw a line of best fit through the points.

3. Plot the millivolt readings of the samples on the standards line and read the corresponding concentrations from the logarithmic scale. These are the concentrations of the nitrate of the samples.

Calcium, Sodium, and Magnesium (Ca^{2+}, Mg^{2+}, Na^+)

Method: Calcium: Flame emission spectrophotometry

Sodium, Magnesium: Atomic absorption spectrophotome-

try (Skougstad and others, 1970; Perkin-Elmers, 1973A)

a.	Nitrate Standard Concentration	Millevolts
	10	136
	100	96
	1000	50
	10000	1
	Sample	119
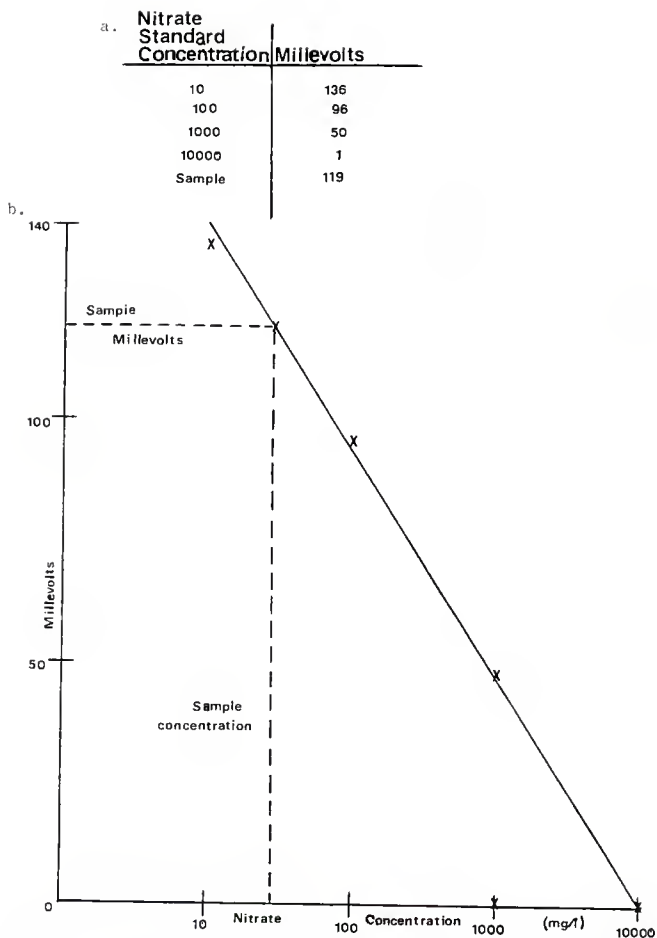
Figure 37. The steps used in calculating the nitrate concentrations a. A table of the mill volt readings of the standards and samples. b. The plot of the data in part a.

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Apparatus: Perkin-Elmer 305-B Spectrophotometer, .500,

.100, and .050 ml micropipettes.

$LaCl_3 \cdot H_2O$ (Lanthanum Chloride), NaCl (Sodium

Chloride), 1000 mg/L calcium, magnesium, and

sodium standards.

Procedure:
1. Prepare a solution containing 1000 mg/l lanthanum by diluting 5.371 g of lanthanum chloride to 1.0 liter

   using D-D water.
2. Add .500 ml of sample water and add 25 ml of the lanthanum solution to a 50 ml volumetric flask and dilute the

   solution to 50 ml with D-D water.
3. Mix thoroughly and pour the prepared sample into a plastic beaker and seal the beaker with cellophane and a

   rubber band.
4. Prepare a standard containing 4.0 mg/l calcium, 4.0 mg/l magnesium, and 2.0 mg/l sodium by pipetting .400 ml of

   calcium standard, .400 ml of magnesium, and .200 ml of sodium standard into the same 100 ml volumetric flask and adding 25 ml lanthanum solution to the flask and

   diluting it to 100 ml.
5. Mix thoroughly and pour into a plastic beaker. Seal the

   beaker with cellophane and a rubber binder.
6. Pipette .100 ml each of calcium and magnesium standards into a 100 ml flask and pipette .040 ml of sodium standard into the volumetric flask. Add 25 ml of lanthanum

   solution to the flask and fill with D-D water.
7. Mix thoroughly and pour it into a plastic beaker. Seal

   the beaker tightly with cellophane and a rubber binder.
8. Pipette .300 ml of calcium and magnesium standard and .050 ml of sodium solution into a 100 ml volumetric flask. Add 50 ml of lanthanum solution to the flask and

   fill it with D-D water.
9. Mix thoroughly and pour it into a plastic beaker. Seal

   the beaker tightly with cellophane and a rubber binder.
10. Pipette .200 ml of calcium and magnesium standard and .100 ml of sodium solution into a 100 ml volumetric flask. Add 50 ml of lanthanum solution to the flask and

    fill it with D-D water.

11. Mix thoroughly and pour it into a plastic beaker. Seal the beaker tightly with cellophane and a rubber binder.

12. Analyse the calcium using atomic emission spectrophotometry and analyse the sodium and magnesium using atomic absorption spectrophotometry. Follow the instructions given by the Perkin-Elmer Corporation (1973B) and use the instrument settings listed in Table 10.

Calculations:

1. The absorbance values of the standards should be plotted against the standards concentrations using arithmetic graph paper (Fig. 38). The plotted points should form a straight line.

2. Plot the absorbance values of the samples against the line and read off the concentration values of the samples off the concentration scale.

3. Multiply the concentration values by 100 to obtain the true concentration of the element in the sample water.

Standard Deviation: Calcium, 4.0 percent.

Magnesium, 15 percent.

Sodium, 8.0 percent (Skougstad and others, 1970)

Potassium and Strontium (Sr^{2+}, K^+)

Method: Potassium: Flame emission Spectrophotography

Strontium: Atomic Absorption Spectrophotometry

(Skougstad and others, 1970; Perkin-Elmers, 1973A).

Apparatus: Perkin-Elmer 305B Atomic Absorption Spectrophotometer, .100, .050, and .020 ml micropipette.

Reagents: $LaCl_3 \cdot 7H_2O$ (Lanthanum Chloride), 1,000 mg/l potassium standard and 1,000 mg/l strontium standard.

Procedure:

1. Prepare a solution containing 2,000 mg/l lanthanum by diluting 5.371 g of lanthanum chloride to 1.0 liter using D-D water.

2. Pipette 10 ml of sample water and 10 ml of 2,000 mg/l lanthanum into the same plastic beaker.

3. Seal the beaker tightly with cellophane and a rubber

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Table 10. Instrument settings used for the atomic absorption spectro-
photometric analyses of cations in the waters of the area (Perkin-
Elmer, 1970).

Element	Wavelength (nm.)	Slit	Range	Function	Burner Width	Burner Height	EM Chopper	Acetylene Pres.	Air Pres.
Calcium	208.6	4	VS	EMS.	2.5	6.0	ON	4.8	5.0
Magnesium	285.2	4	UV	ABS.	2.5	7.5	OFF	5.0	5.3
Sodium	291.6	4	VS	ABS.	2.5	6.0	OFF	6.0	5.0
Potassium	382.4	4	VS	EMS.	2.5	6.0	ON	4.7	5.3
Strontium	227.7	4	VS	ABS.	2.5	7.2	OFF	4.7	5.0
Iron	248.3	3	VS	EMS.	2.5	7.0	OFF	5.0	5.0
Manganese	279.5	4	UV	ABS.	2.5	8.2	OFF	5.0	6.0

Note: Acetylene and air pressure are given as pounds per square inch.
Note: Potassium is analysed using a red filter.
Note: Burner height and width are given in inches.

b.

Calcium Standard (mg/l) Concentration	Absorption
Blank	5.0
1.0	10.0
2.0	15.0
3.0	20.0
4.0	25.0
Sample	12.5
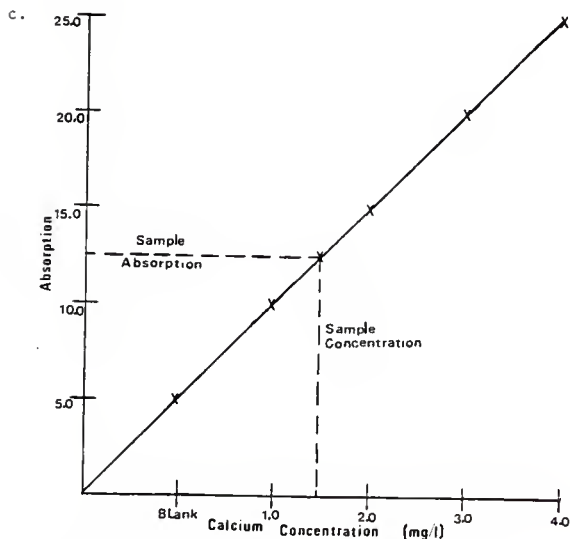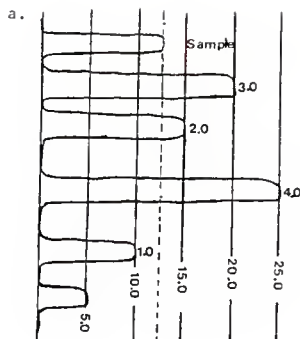
d. Calcium concentration = sample concentration (dilution).

Fig. 38. The steps in calculating the calcium concentrations:
a. The strip chart of the data. b. Table of the strip
chart data. c. Plot of the strip chart data. d. Calcu-
lations of the actual calcium concentration of the sam-
ples.

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band. Pipette .050 ml each of 1,000 mg/l potassium stan-
dard and strontium standard into a 25 ml volumetric
flask and dilute to 25 ml with D-D water.

4. Pipette .070 ml each of 1,000 mg/l potassium and stron-
tium standard into a 25 ml volumetric flask and dilute

to 25 ml with D-D water.
5. Pipette .100 ml each of 1,000 mg/l potassium and stron-
tium standard into a 25 ml volumetric flask and dilute

to 25 ml with D-D water.
6. Prepare 1.0, 1.4, and 2.0 mg/l standards by Pipetting
10 ml of each of the solutions prepared above into plas-
tic beakers and adding 10 ml of 2,000 mg/l lanthanum to

the beakers.
7. Seal the beakers tightly with cellophane and a rubber
band. Run the samples and the solutions on the atomic
absorption spectrophotometer using the instructions in
the Perkin-Elmer Manual "Instructions Model 305B Atomic
absorption Spectrophotometer" (1973B), and using the

instrument settings listed in Table 10.

Calculations:
1. Plot the standards on arithmetic graph paper with the
absorption on one axis and the concentration on the
other axis (Fig.33). Draw a straight line through the
points formed by the standards.

2. Plot the absorption of the samples on the sample line
and read the concentration of the sample solution from

the cocentration axis.
3. Multiply this figure by two to obtain the samples true

concentration.
Standard Deviation: Potassium, 15 percent
                    Strontium, 11 percent (Skougstad and

                    others, 1970)

Iron and Manganese (Fe $^{2+}$, Mn $^{2+}$)

Method: Atomic absorption spectrophotometry (Skougstad and

       others, 1970: Perkin-Elmers, 1973A)

Apparatus: Perkin-Elmer 305B atomic absorption spectropho-

          tometer

Reagents: Concentrated HNO_3 (nitric acid), NaCl (Sodium

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chloride)

Procedure:
1. Prepare a 200 mg/l solution of sodium by dissolving
   .5084 g of sodium chloride and diluting it to a volume

   of 1.0 liter with D-D water.
2. Place 50 ml of sample and 50 ml of sodium solution into

   a glass beaker.
3. Pipette .020 ml of 1,000 mg/l iron standard and manga-
   nese standard into a 50 ml volumetric flask and fill

   with D-D water.
4. Pipette .040 ml of 1,000 mg/l iron standard and manga-
   nese standard into a 50 ml volumetric flask and fill

   with D-D water.
5. Pipette .050 ml of 1,000 mg/l iron standard and manga-
   nese standard into a 50 ml volumetric flask and fill

   with D-D water.
6. Pour each of the prepared standards into a glass beaker

   and add 50 ml of sodium solution to each of them.
7. Pour 50 ml of D-D water into a glass beaker and add 50

   ml of sodium solution to it.
8. Pipette 1.0 ml of nitric acid in to each of the samples
   and standards. Pipette .100 ml of the hydrocloric acid
   used to preserve the samples to the blank and the

   standards.
9. Evaporate the samples, blank and standards to a volume

   of less than 10 ml using a hotplate.
10. Pour the samples into a 10 ml volumetric flask and use
    several small rinses to clean the beaker. Use the water
    from the rinses to dilute the sample to a volume of 10

    ml.
11. Pour into a plastic beaker and seal tightly with cello-

    phane and a rubber band.
12. Run the samples using the instructions found in the
    Perkin-Elmer manual "Instructions Model 305B Atomic
    Absorption Spectrophotometer" (1973B), and using the

    instr ment settings listed in Table 10.

Calculations:
1. Plot the standards on arithmetic graph paper with the
   concentration on one axis and the absorption on the
   second axis and draw a line through the points (Fig.
   33).

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2. Plot the position of the absorption of the samples on
   the standards line and read off the corresponding con-

   centration value for the samples.
3. Multiply the values by five to obtain the concentration

   of iron or manganese in the sample water.
Standard Deviation: Iron, 8.0 percent
                    Manganese, 14.0 percent (Skougstad

                    and others, 1970).

APPENDIX   V

Appendix V. The results of the three sets of analyses of water samples from the third series of samples.

Part 1. The results of the analyses conducted for the study.

well location	well owner	sample date	temp. (C)	spec. cond	pH	Ca	Mg	Na	K	Sr	Fe	Mn	HCO	SO	Cl	NO
8 9 35 cdd#3	Blanka	3/6/86	11.3	998	7.0	93	33	74	2.7	26	DL	DL	403	222	13	8
9 8 26dbd#3	Thiele	3/4/86	14.8	724	7.0	87	17	11	1.0	1.0	DL	DL	355	32	12	54
9 9 4ddd#3	Ebert	3/7/86	11.8	6777	7.2	119	22	24	2.1	1.0	3.1	1.2	393	35	13	6
9 9 15dbc#3	Patnode	3/6/86	7.4	401	7.1	72	11	23	2.3	.25	.40	.70	292	19	2.0	DL
9 9 24add#3	Warnow	3/6/86	13.7	241	6.8	39	3.4	15	1.4	.10	DL	DL	172	2.0	2.0	3
9 9 31baa#3	D. Ubel	3/7/86	7.3	357	6.9	50	7.8	16	2.4	.35	DL	DL	250	2.0	1.0	6
9 9 33adb#3	Piper	3/4/86	13.8	86	6.7	15	2.0	9.0	.1	.10	DL	DL	101	2.0	1.0	5
9 10 16cac#3	Umschied	3/6/86	14.0	973	6.7	117	25	44	1.6	.29	.13	.14	239	92	69	180
9 10 28bda#3	Bammes	3/5/86	10.7	282	6.5	40	4.9	18	1.5	.15	DL	DL	138	3.0	10	7
9 10 32cbc#3	Seele	3/5/86	14.0	430	7.1	49	4.8	31	1.7	.19	DL	DL	224	2.0	12	5
10 8 12cab#3	Kern	3/4/86	13.8	596	7.1	81	9.1	25	3.2	.26	DL	DL	321	14	6	1
10 9 3ddc#3	Altwegg	3/10/86	14.0	386	7.2	61	8.0	6.5	.8	.16	.14	DL	223	14	2	8
10 9 5ddc#3	Carver	3/10/86	14.3	383	6.8	49	6.6	13	2.9	.21	DL	DL	108	14	13	62
10 9 15bbb#3	Oppen.	3/7/86	11.0	1496	6.8	302	39	20	2.9	.33	DL	DL	400	612	12	8
10 10 6bdd#3	Hildreth	3/4/86	13.7	253	6.4	25	4.0	21	1.0	.10	DL	DL	105	8	8	20
10 10 10abb#3	Fairchild	3/5/86	14.0	1250	7.0	175	24	31	25	.52	DL	DL	324	121	51	220

Note: the specific conductivity is expressed in micromhos per centimeter at 25 C.
Note: the concentrations are expressed in milligrams per liter.

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Part 2. The results of the analyses conducted by the Kansas
        State Geologic Survey.

well location	well owner	spec. cond.	pH	Ca	Mg	Na	K	HCO	SO	Cl	NO
10 8 12cab#3	Kern	505	7.42	76	8.7	25	3.5	295	13	5.4	3.8
10 9 3ddc#3	Altwegg	343	7.60	58	6.9	7.7	0.9	193	8.8	1.5	5.7
10 9 5ddc#3	Carver	350	7.20	47	5.9	13	2.2	110	14	12	56

Part 3. The results of the analyses conducted by the Kansas
State University Chemical Emission Spectroscopy Laboratory.

well Location	well owner	Ca	Mg	Na	K	Fe	Mn
8 9 35cdd#3	O. Blanka	92	29	75	2.9	076	.018
9 8 26dbd#3	E.H. Thiele	76	17	9.5	1.0	0L	0L
9 9 4ddd#3	F. Ebert	120	24	28	2.3	3.1	1.2
9 9 15dbc#3	W. Patnode	84	14	23	3.4	.40	.73
9 9 24add#3	F. Warnow	37	3.6	17	2.0	0L	0L
9 9 30baa#3	O. Ubel	51	7.9	32	2.9	0L	0L
9 9 33adb#3	K. Piper	16	1.9	9.5	1.6	0L	0L
9 10 16cac#3	O. Umschied	120	27	76	1.5	.13	.14
9 10 28bda#3	L. Bammes	40	5.1	17	2	0L	0L
9 10 32cbc#3	A. Seele	51	5.2	34	1.9	0L	0L
10 8 12cab#3	K. Kern	85	9.1	25	4.0	0L	.041
10 9 3ddc#3	O. Altwegg	62	6.1	25	4.0	.14	.028
10 9 5ddc#3	L. Carver	52	6.4	15	3.0	.035	0L
10 9 15bbb#3	Oppenlander	310	43	24	4.1	.039	.018
10 10 6bdd#3	J. Hildreth	24	4.7	21	1.3	0L	0L
10 10 10abb#3	Fairchild	180	25	33	25	0L	0L

Note: The concentrations are expressed in milligrams per liter.
Note: the specific conductivity is expressed in micromhos.

A P P E N D I X   VI

Appendix VI. The results of the mass balance of the anions and cations for the waters of sample seires two, three, and four.

Sample Series Two

Well ocation	Owner	Sum of the Cations	Sum of the Anions	Percent Difference
8 9 35cdd	O. Blanka	12.6289	11.5435	+4.5
9 8 26dbd	E.H. Thiele	8.1057	8.1776	-.44
9 9 4ddd	F. Ebert	8.8987	8.9310	-.18
9 9 15dbc	W. Patnode	5.4470	5.3605	+.80
9 9 24add	F. Warnow	2.8129	2.8662	-.94
9 9 30baa	D. Ubel	4.1662	4.4710	-3.5
9 9 33adb	K. Piper	1.5868	1.7052	-3.6
9 10 16cac	D. Umschied	13.7252	15.4462	-5.9
9 10 28bda	L. Bammes			
9 10 32cbc	A. Seele	4.3722	4.5855	-2.4
10 8 12cab	K. Kern	6.6383	6.5409	+.74
10 9 3ddc	D. Altwegg	4.0358	4.1154	-.98
10 9 5ddc	L. Carver	2.6127	3.4363	-14.
10 9 15bbb	Oppenlander	18.2286	19.1972	-2.6
10 10 6bdd	J. Hildreth	2.8548	2.8450	+.17
10 10 10abb	R. Fairchild	10.0007	10.5214	-2.5

Sample Series Three

Well Location	Owner	Sum of the Cations	Sum of the Anions	Percent Difference
8 9 35cdd	O. Blanka	11.2377	11.7229	-2.1
9 8 26dbd	E.H. Thiele	6.2671	7.6452	-9.9
9 9 4ddd	F. Ebert	9.0240	10.1100	-5.7
9 9 15dbc	W. Patnode	5.6033	5.2795	+3.0
9 9 24add	F. Warnow	2.9165	2.9655	-.83
9 9 30baa	D. Ubel	3.9023	4.2314	-4.0
9 9 33adb	K. Piper	1.3350	1.7241	-13.
9 10 16cac	D. Umschied	9.8516	10.4706	-3.0
9 10 28bda	L. Bammes	3.2248	2.7931	+8.5
9 10 32cbc	A. Seele	4.2318	3.7931	+5.5
10 8 12cab	K. Kern	5.9661	5.8833	+1.4
10 9 3ddc	D. Altwegg	4.1340	4.1319	+.03
10 9 5ddc	L. Carver	3.6327	3.3956	-1.4
10 9 15bbb	Oppenlander	19.2317	19.7000	-1.2

Note: The sum of the anions and the sum of the cations are expressed as milligram equivalent weights.

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Well Location	Owner	Sum of the Cations	Sum of the Anions	Percent Difference
10 10 6bdd	J. Hildreth	2.5181	2.5967	-1.5
10 10 10abb	R. Fairchild	12.7074	12.8168	-.4

Sample Series Four

Well Location	Owner	Sum of the Cations	Sum of the Anions	Percent Difference
9 9 4ddd	F. Ebert	8.7943	7.8458	+5.7
9 9 24add	F. Warnow	2.5953	2.8608	-4.9
9 9 30baa	D. Ubel	4.3076	5.0416	-7.9
9 9 33adb	K. Piper	1.6981	2.2026	-13.
9 10 16cac	D. Umschied	11.5137	10.2208	+5.9
10 9 5ddc	L. Carver	3.2631	3.3239	-1.0
10 9 15bbb	Oppenlander	18.4932	19.1086	-1.6
10 10 10abb	R. Fairchild	10.8361	9.2368	+2.7

Note: The sum of the anions and the sum of the cations are expressed as milligram equivalent weights.

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APPENDIX   VII

Appendix VII. Well Logs of Sampled Wells and Wells Used in
Stratigraphic Correlation Diagrams.

The following pages contain the well logs of water
wells submitted to the Kansas State Department of Health
and Enviorment by water well drillers in Kansas. The wells
in this appendix were used to construct the stratigraphic
correlation diagrams in the thesis or had their water sam-
pled during the investigation. The wells are listed in the
order of their land office number starting with T. 8S. R.
8E. 1 and ending with T.10S. R.10E. 19. The elevations of
the wells were determined to the nearest 5 ft using the St.
George, Flush, Wamego and Louisville 7.5 minute topographi-
cal quadrangle maps of the United States Geological Survey.
The depths and thicknesses in the well logs below are given
in feet because the original well logs were made in the
english units of measurements.

Water well located T.9E. R.8S. sec. 25dcd and owned by Da-
vid Irwin. Drilled 10/12/78 by Strader Drilling Co. Water
level 87 ft below the surface and the wells yield is .5
gallons per minute. Well depth, 87 ft below the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
clay	0	22
Permian		
limestone	22	3
blue shale	25	3
blue limestone	28	22
blue shale	50	16
red shale	66	4
blue shale	70	

Water well located T.9E. R.8S. sec. 26aba and owned by Mike
Briggs. Drilled 9/2/83 by Strader Drilling Co. Water level
78 ft below the surface and the well yield 15 gallons per
minute. Well depth, 160 ft below the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	3
Permian		
yellow shale	3	14
yellow limestone	17	10
yellow shale	27	14
yellow limestone	51	4
yellow shale	55	22
yellow limestone	77	4
gray shale	79	12

Description of material	depth from surface (ft)	thickness (ft)
gray limestone	91	2
gray shale	93	18
black shale	111	4
gray shale	115	4
gray limestone	119	6
gray shale	125	16
red shale	141	9
Pennsylvanian (?)		
gray limestone	150	2
gray shale	152	

Water well located T.9E. R.8S. sec. 26dbd and owned by E.H. Thiele. Drilled by Strader Drilling Co. Water level 19 ft. below the surface and the well depth is 45 ft. below the surface. The well yield is 20 gallon per minute.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	3
yellow clay	3	12
sand and gravel	15	3
Pennsylvanian		
blue shale	18	

Water well located T.9E. R.8S. sec. 27cad and owned by Coffman Plumbing. Drilled 10/27/77 by Strader Drilling Co. Water level 50 ft below the surface and the well yield is 30 gallons per minute. Well depth, 240 ft below the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	3
brown clay	3	15
Permian		
gray limestone	18	17
gray shale	35	12
yellow limestone	47	3
gray shale	50	15
loose gray limestone	65	1
gray shale	66	

Water well located T.9E. R.9S. sec. 4ddd and owned by Fred Ebert. Drilled 8/11/82 by Strader Drilling Co. Water level 20 ft below the surface and the well yields 43 gallons per minute. The well depth is 38 ft below the surface.

Description of material	depth from surface (ft)	thickness (ft)

```
Quaternary
    top soil                        0           3
    brown clay                      3          12
    chert gravel                   15           6
    brown clay                     21          11
    chert gravel                   32           3
Pennsylvanian
    shale                          35
```

Water well located T.9E. R.9S. sec. 11daa and owned by Eric
Bartholomew. Drilled 3/21/78 by Strader Drilling Co. Water
level 25 ft below the surface. The well depth is 100 ft
below the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	4
clay	4	19
coarse sand and gravel	23	6
Pennsylvanian		
shale and limestone	29	

Water well located T.9E. R.9S. sec. 15dbc and owned by
William Patnode. Drilled 5/14/80 by Strader Drilling Co.
Water level 37 ft below the surface and the well yields 40
gallons per minute. The well depth is 80 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	6
brown and blue clays	6	31
fine sand, coarse sand, and chert gravel	37	2
Pennsylvanian (?)		
grey shale	39	

Water well located T.9E. R.9S. sec. 24add and owned by Fred
Warnow. Drilled 3/21/78 by Strader Drilling Co. Water level
25 ft below the surface and the well yields 20 gallons per
minute. The well depth is 80 ft below the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	4
clay	4	41
gravel	45	

Water well located T.9E. R.9S. sec. 26cdd and owned by
Bruce Flanagan. Drilled 3/21/78 by Strader Drilling Co.
Water level 70 ft below the surface and the well yields 30
gallons per minute. The well depth is 113 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	5
silty clay	5	80
fine and coarse sand	80	

Water well located T.9E. R.9S. sec. 28dcd and owned by
Doyle Danker. Drilled 7/30/76 by Strader Drilling Co. Water
level 80 ft below the surface and the well yields 50 gal-
lons per minute. The well depth is 135 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	4
brown sandy clay	4	71
fine sand	75	50
coarse sand	125	10

Water well located T.9E. R.9S. sec. 29cdc and owned by
Pottawatomie County Rural Water District Number One. Dril-
led 1/27/75 by Strader Drilling Co. Water level 84 ft below
the surface and the well yields 300 gallons per minute. The
well depth is 178 ft below the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	2
brown clay	2	6
red clay	8	12
fine sand	20	45
clay	65	32
fine sand	97	13
clay	110	22
fine sand with clay	132	13
fine to coarse clayey sand	145	8
coarse sand and gravel	153	20
sandy clay	173	3
gravel and coarse sand	176	1

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Permian (?)
        blue shale   177   1

Water well located T.9E. R.9S. sec. 31baa and owned by
Delbert Ubel. Drilled 6/12/79 by Strader Drilling Co. Water
level 83 ft below the surface and the well yields 20 gal-
lons per minute. The well depth is 125 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	2
clay	2	10
fine sand	12	3
tan clay	15	4
very fine sand	19	13
tan clay	32	7
clayey sand	39	31
fine sand	70	4
tan clay	74	17
very fine sand	91	7
blue clay	98	9
gravel	107	7
Permian		
blue shale	114	

Water well located T.9E. R.9S. sec. 31cbc and owned by
Donald Lamb. Drilled 12/1/75 by Strader Drilling Co. Water
level 38 ft below the surface and the well yields 6 gallons
per minute. The well depth is 240 ft below the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	4
yellow clay	4	19
Permian		
blue shale	23	17
yellow limestone	40	5
gray shale	45	30
gray limestone	75	4
blue shale	79	

Water well located T.9E. R.9S. sec. 33adb and owned by
Kenneth Piper. Drilled 8/1/75 by Strader Drilling Co. The
well yields 50 gallons per minute. The well depth is 144 ft
below the surface.

depth from   thickness

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	5
yellow sandy clay	5	95
fine sand	100	44
Permian (?)		
limestone	144	

Water well located T.9E. R.9S. sec. 34aab and owned by
Julius Lintz. Drilled 8/18/78 by Strader Drilling Co. Water
level 50 ft below the surface and the well yields 50 gal-
lons per minute. The well depth is 150 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	4
brown clay	4	36
silty fine sand	40	

Water well located T.9E. R.9S. sec. 36bcc and owned by
Larry Hudson. Drilled 3/30/81 by Strader Drilling Co. Water
level 50 ft below the surface. The well depth is 68 ft
below the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	1
yellow silty clay	1	59
yellow medium sand	60	

Water well located T.9E. R.10S. sec. 3ddd and owned by
Lloyd Blanka. Drilled 3/21/78 by Strader Drilling Co. Water
level 25 ft below the surface and the well yields 30 gal-
lons per minute. The well depth is 80 ft below the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	6
sandy clay	6	74
gravel	80	

Water well located T.9E. R.10S. sec. 6aab and owned by
Darold Brunkow. Drilled 2/14/72 by Strader Drilling Co.

Water level 28 ft below the surface and the well yields 6
gallons per minute. The well depth is 60 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	6
yellow clay	6	22
rock	28	1
Pennsylvanian (?)		
blue limestone	29	6
white and blue shale	35	

Water well located T.9E. R.10S. sec. 14aba and owned by
Lyle Whearty. Drilled by Strader Drilling Co. Water level
15 ft below the surface and the well yields 4 gallons per
minute. The well depth is 60 ft below the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	3
mud	3	40
Permian		
brown limestone	43	2
shale	45	15
limestone	60	

Water well located T.9E. R.10S. sec. 16cac and owned by
Dennis Umschied. Drilled 7/29/80 by Strader Drilling Co.
Water level 30 ft below the surface. The well depth is 63
ft below the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	4
brown clay	4	21
yellow clay	25	13
Pennsylvanian (?)		
limestone	38	10
shale	48	

Water well located T.9E. R.10S. sec. 28bda and owned by
Larry Bammes. Drilled 8/14/79 by Strader Drilling Co. The
water level is 15 ft below the surface and the well yields
4 gallons per minute. The well depth is 58 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)

Quaternary
```
    top soil                          0          6
    brown clay                        6          9
    fine and coarse sand             15          5
Permian (?)
    gray shale                       20
```

Water well located T.9S. R.10S. sec. 29dcc and owned by
Pottawatomie County. Drilled by Strader Drilling Co. The
well depth is 73 ft below the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	8
fine sand	8	15
silt and tan clay	23	47
Pennsylvanian (?)		
blue shale	70	

Water well located T.9E. R.10S. sec. 31aaa and owned by
John Winterman. Drilled 7/3/78 by Strader Drilling Co.
Water level 65 ft below the surface and the well yields 30
gallons per minute. The well depth is 90 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	4
brown clay	4	59
fine sand	63	15
coarse sand and gravel	78	7
Pennsylvanian (?)		
shale	85	

Water well located T.9E. R.10S. sec. 32cbc and owned by Al
Seele. Drilled 1/28/75 by Strader Drilling Co. The water
level is 65 ft below the surface and the well yields 30
gallons per minute. The well depth is 120 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	5
yellow clay	5	35
brown clay	40	30
coarse and fine sand	70	14
coarse sand and medium		

gravel	84	12
Pennsylvanian (?)		
blue shale	96	

Water well located T.9E. R.10S. sec. 33bbd and owned by
Irwin Duncan. Drilled 1/29/75 by Strader Drilling Co. The
well yields 30 gallons per minute. The well depth is 124 ft
below the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	5
yellow clay	5	35
blue clay	40	17
brown clay	57	23
coarse sand	80	10
coarse sand and		
small gravel	90	

Water well located T.9E. R.10S. sec. 35bbc and owned by
Everett Boyce. Drilled 3/27/76 by Strader Drilling Co. The
water level is 29 ft below the surface and the well yields
3000 gallons per minute. The well depth is 83 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	3
soft brown clay	3	28
fine gray sand	31	1
medium gray gravel	32	4
medium gray gravel		
and clay	36	1
gray clay	37	9
gray gravel and clay	46	1
medium gray gravel	47	5
large gray gravel	52	30
large gray gravel and		
clay	82	1
gray clay	83	

Water well located T.9E. R.10S. sec. 36cda and owned by
Leroy Miller. Drilled 6/22/78 by Strader Drilling Co. The
water level is 28 ft below the surface and the well yields
1500 gallons per minute. The well depth is 57 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)

164

```
Quaternary
    top soil                            0          3
    light brown clay                    3         23
    small brown gravel                 26          3
    medium brown gravel                29          1
    small brown gravel and
        sand                           30          1
    gray clay                          31          3
    fine brown sand                    34          2
    small brown gravel and
     gray clay                         36          5
    medium brown gravel                41          5
    medium and large brown
        gravel                         46          2
    medium and large green
        gravel                         48          3
    small and medium green
        gravel                         51          2
    medium and large green
        gravel                         53
```

Water well located T.9E. R.11S. sec. 6bcc and owned by
Robert Peterson. Drilled 1/8/81 by Strader Drilling Co. The
water level is 35 ft below the surface and the wells yield
is 30 gallons per minute. The well depth is 58 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	6
brown clay	6	29
fine and coarse sand	35	14
Permian		
gray shale	49	

Water well located T.10S. R.8S. sec. 2bcd and owned by T.
Ubel Drilled by Strader Drilling Co. The water table is 57
ft below the surface and the well yields 10 gallons per
minute. The well depth is 82 ft below the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	5
Brown clay	5	35
fine sand	40	37
gravel and fine sand	77	3
Permian		
shale	80	

Water well located T.10E. R.8S. sec. 12cab and owned by
Kenneth Kern. Drilled 6/22/78 by Strader Drilling Co. The
water level is 18 ft below the surface and the wells yield
is 50 gallons per minute. The well depth is 39 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	6
brown clay	6	12
fine sand, coarse sand		
and medium gravel	18	14
green gravel	32	

Water well located T.10E. R.9S. sec. 3dbb and owned by
Michael Quinn. Drilled 9/30/30 by Strader Drilling Co. The
water level is 80 ft below the surface and the well yields
50 gallons per minute. The well depth is 125 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
sandy silt	0	7
sandy fine silt	7	73
fine sand	80	45
Pennsylvanian		
limestone	125	

Water well located T.10E. R.9S. sec. 3ddc and owned by
David Altwegg. Drilled 4/28/82 by Strader Drilling Co. The
water level is 70 ft below the surface and the well yields
5 gallons per minute. The well depth is 120 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	3
fine sand	3	34
brown clay	37	4
Pennsylvanian		
yellow shale	41	12
gray limestone	53	2
gray shale	55	30
gray limestone	85	20
gray sandstone	105	

Water well located T.10S. R.9S. sec. 5cdb and owned by C.L.
Woodard. Drilled 10/11/71 by Strader Drilling Co. The well
yields 20 gallons per minute. The well depth is 94 ft below
the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	4
Brown clay	4	16
fine sand	20	14
tan clay	34	4
fine sand	38	19
tan clay	57	2
fine sand	59	15
clay	74	2
fine sand	94	

Water well located T.10E. R.9S. sec. 5ddc and owned by Levi
Carver. Drilled 8/8/83 by Strader Drilling Co. The water
level is 47 ft below the surface and the well yields 30
gallons per minute. The well depth is 84 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
fine sand	0	10
brown clay	10	42
fine sand	52	4
brown clay	56	12
fine sand, coarse sand,		
and medium gravel	68	10
chert gravel	78	1
Pennsylvanian		
limestone	79	2
shale	81	

Water well located T.10E. R.9S. sec. 6ada and owned by Carl
Meyer. Drilled 12/6/78 by Strader Drilling Co. The water
level is 60 ft below the surface and the well yields 100
gallons per minute. The well depth is 122 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	6
brown clay	6	54
fine silty sand	60	35
fine and coarse sand	95	24
Permian		

gray shale   119

Water well located T.10E. R.9S. sec. 7baa and owned by
Dudley Torrey. Drilled 3/17/82 by Strader Drilling Co. The
water level is 65 ft below the surface and the well yields
10 gallons per minute. The well depth is 110 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	2
brown clay	2	5
sandy clay	7	20
brown clay	27	6
fine sand	33	39
clay	72	7
medium fine sand	79	6
light gray clay	85	5
fine sand	90	10
medium fine sand	100	2
Permian		
gray shale	102	

Water well located T.10E. R.9S. sec. 9cad and owned by City
of Wamego. Drilled 3/30/81 by Strader Drilling Co. The
water level is 79 ft below the surface and the well yields
35 gallons per minute. The well depth is 103 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	7
sandy silt	7	38
sandy clay	45	39
fine and coarse sand	74	10
coarse sand and medium to coarse gravel	84	14
Pennsylvanian		
gray shale	98	

Water well located T.10E. R.9S. sec. 14dac and owned by
Robert Shaw. Drilled 6/2/75 by Strader Drilling Co. The
water level is 18 ft below the surface and the well yield
is 100 gallons per minute. The well depth is 47 ft below
the surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		

```
top soil                      0        15
clay                         15         3
sand                         18        10
sand and gravel              28         9
gravel                       37
```

Water well located T.10E. R.9S. sec. 15abd and owned by
Dennis Drouillard. Drilled 6/13/80 by Strader Drilling Co.
The water level is 58 ft below the surface. The well depth
is 100 ft below the surface.

```
                            depth from   thickness
    Description of material  surface (ft)    (ft)

Quaternary
    fine sand                    0          32
    sandy clay                  32          45
Pennsylvanian
    gray loose limestone        77           1
    gray shale                  78
```

Water well located T.10E. R.9S. sec. 15bbb and owned by
Marvin Oppenlander. Drilled 8/21/81 by Strader Drilling Co.
The water level is 20 ft below the surface and the well
yields 5 gallons per minute. The well depth is 60 ft below
the surface.

```
                            depth from   thickness
    Description of material  surface (ft)    (ft)

Quaternary
    top soil                     0           6
    silty clay                   6           8
    yellow, clean, fine sand    14          12
    chert gravel                26           2
Pennsylvanian
    gray shale                  28           8
    gray limestone              36           5
    gray shale                  41          10
    gray limestone              51           2
    gray shale                  53
```

Water well located T.10E. R.9S. sec. 17cac and owned by
Darold Doperalski. Drilled 4/11/79 by Strader Drilling Co.
The water level is 15 ft below the surface and the well
yields 50 gallons per minute. The well depth is 60 ft below
the surface.

```
                            depth from   thickness
    Description of material  surface (ft)    (ft)

Quaternary
    top soil                     0           6
    brown clay                   6          12
    fine sand, coarse sand
```

169

and gravel   18

Water well located T.10E. R.10S. sec. 1bda and owned by
Lyle Slyvestor. Drilled 6/2/75 by Strader Drilling Co. The
water level is 18   ft below the surface and the well yields
1200 gallons per minute. The well depth is 47 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil and clay	0	14
brown gravel	14	12
sand	26	2
gravel	28	4
sand	32	4
gravel	36	5
sand	41	2
gravel	43	5
Permian		
clay and shale	48	6
shale	54	

Water well located T.10E. R.10S. sec. 6bda and owned by J.
Hildreth. Drilled 3/2/81 by Strader Drilling Co. The water
level is 60 ft below the surface and the well yields 8
gallons per minute. The well depth is 90 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	6
brown clay	6	34
fine sand	40	9
brown clay	49	19
fine and coarse sand	68	10
Pennsylvanian		
gray shale	78	

Water well located T.10E. R.10S. sec. 10abb and owned by
Ray Fairchild. Drilled 9/27/78 by Strader Drilling Co. The
water level is 25 ft below the surface and the well yields
500 gallons per minute. The well depth is 47 ft below the
surface.

Description of material	depth from surface (ft)	thickness (ft)
Quaternary		
top soil	0	6
brown clay	6	19
fine sand	25	11

```
fine sand, coarse sand,
   and medium gravel          36
```

A P P E N D I X   VIII

Appendix VIII. The depth to the bedrock, the depth to the water table, the aquifer type, the surface elevation, the total depth, and the yield of the wells used to construct the subsurface contour map of the buried valley.

Well location	well owner	well depth	surface elev.	depth to bedrock	depth to water	aquifer type	well yield (GPM)
8 9 35cdd O. Blanka		92	1120		66	BR	4
9 8 2ddd D. Kunidger		240	1250	18	172	BR	2
9 8 11bdd C. Bellinger		62	1250	10	30	BR	
9 8 25dcd D. Irwin		87	1230	22	24	BR	.5
9 8 26aba M. Briggs		160	1180	3	78	BR	15
9 8 26bcc M. Dikeman		140	1250	75	95	BR	10
9 8 26dbd E. Thiele		45	1140	18	19	BR	20
9 8 27cad Coffman		80	1210	18	50	BR	30
9 9 4ddc F. Ebert		60	1040	46	40	UA	15
9 9 4ddd F. Ebert		38	1045	35	20	UA	12
9 9 7ddd B. Heptig		60	1140	13	12	Q	8
9 9 11daa Bartholomew						UA	
9 9 15dbc W. Patnode		80	1070	39	37	BV	20
9 9 24add F. Warnow		80	1040	50	45	BV	6
9 9 25cdd M. Walsh		115	1080	114	17	Q	80
9 9 26cdd B. Flannigan		113	1135	113+	70	BV	30
9 9 28bdd Weisbender		160	1180	160+	73	BV	100
9 9 28dcd D. Danker		135	1165	135+	80	BV	50
9 9 29cdb PCRWD #1		176	1190	173	86	BV	250
9 9 29cdc PCRWD #1		178	1195	177	84	BV	300
9 9 29dcc PCRWD #1		160	1180	160	82	BV	450
9 9 30aaa G. Burnett		100	1140	100+	80	BV	20
9 9 30aad D. Zarger		100	1150	22	20	BV	5
9 9 30abd D. Westgate		69	1150	0	43	BR	12
9 9 31add J. Sweany		110	1190	108	60	BV	30
9 9 31baa D. Ubel		125	1205	114	85	BV	20
9 9 31cbc D. Lamb		100	1160	23	38	BR	6
9 9 32caa B. Swart		175	1170	173	98	BV	100
9 9 30cbb C. Meyer		150	1200	150	70	BV	50
9 9 33ccc K. Piper		144	1160	144		BV	50
9 9 34aab J. Lintz		120	1090	120+	50	BR	50
9 9 35aaa Cunningham		145	1140	145+	80	BV	80

KA = Kansas River alluvium
UA = upland stream alluvium
BV = buried valley sediments
Q = glacial deposits
BR = bedrock

Note: all elevations and depths are given in feet.
Note: base level is mean sea level.

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Well location	well owner	well depth	surface elev.	depth to bedrock	depth to water	aquifer type	well yield (GPM)
9 9 36bcc	L. Hudson	88		88+	50	BV	
9 10 1ccc	W. Eichman	200	1070	8	60	BR	30
9 10 3ddd	L. Blanka	140	1040	80	80	Q	
9 10 6aab	D. Brunkow	60	1080	29	28	BR	6
9 10 8aca	J. Wilson	78	1080	72	50	BV	15
9 10 9bbb	Pott Co.	73	1090	8	20	BR	30
9 10 12cab	Guilfoil	100	1020	39	39	BR	.5
9 10 14aba	L. Whearty						
9 10 16aaa	L. Yonning	81	1030	28	8	BR	14
9 10 16baa	G. Yonning	80	1035	18	15	Q	1.5
9 10 16cac	D. Umschied	63	1035	38	30	(Q)	
9 10 20baa	L. Brayton	50	1000	43	18	UA	6
9 10 21ccc	T. Picolett	55	1010	55+	30	UA	30
9 10 21dbb	V. Robinson	52	1000	50	20	UA	
9 10 28bda	L. Bammes	58	1020	20	15	UA	4
9 10 28cbd	E. Yeager	78	1040	55		BV	30
9 10 28dbb	C. Godfrey	60	1020	31	18	BV	25
9 10 29dcc	Pott. Co.	82		80		BV	
9 10 30dca	Peddicord	60	970	60+	18	KA	2000
9 10 31aaa	Winterman	90	1060	85	65	BV	30
9 10 32acd	K. Bowen	105	1090	105	65	BV	30
9 10 32bba	Richards	120	1110	115		BV	10
9 10 32bcc	Snapp-Wilson	59	1010	59+	30	BV	20
9 10 32cbc	A. Seele	100	1040	96	65	BV	30
9 10 32dcc	L. Wilson	102	1060	100	59	BV	50
9 10 32dbb	C. Chapman	100	1040	100+	60	BV	30
9 10 32dbd	J. Waters	100	1070	100+	60	BV	50
9 10 33bbd	I. Duncan	124	1060	124+	75	BV	30
9 10 35bbc	E. Boyce	83	980	83+	29	UA	3000
9 10 35dca	Peddicord	59	975	59	27	KA	450
9 10 35dbb	Peddicord	64	975	64	26	KA	675
9 10 36cda	L. Miller	57	970	57+	29	KA	1500
9 10 36dcb	G. Hieger	50	977	50	21	KA	1000
9 11 6bcc	R. Peterson	58	995	49	33	UA	30
9 11 6cba	R. Peterson	58	1000	46	28	UA	30
10 8 2bcd	T. Ubel	82	1095	80	57	BV	6
10 8 11dad	Dehydrator	45	1000	45+	20	KA	100
10 8 12cab	K. Kern	39	1000	39+	18	KA	50
10 8 12dba	C. Cragg	40	1005	32	11	KA	20
10 8 14bcc	Cedar Creek	50	1010	50+	20	KA	20
10 9 2add	Hildreth	92	1200	76	50	BV	12
10 9 3dad	Anderson	116	1180	27	30	BR	10
10 9 3dbb	M. Quinn	125		125	80	BV	50
10 9 3ddc	Altwegg	120	1200	41	70	BR	5
10 9 4aaa	Whitney	75	1140	75+	50	BV	50
10 9 4abb	Hodges	130	1100	120	75	BV	50

Well location	well owner	well depth	surface elev.	depth to bedrock	depth to water	aquifer type	well yield (GPM)
10 9 4bcc	F. Hodges	150	1140	148	75	BV	30
10 9 4cab	M. Crubel	60	1090	60+	30	BV	15
10 9 4ccc	E. Taylor	51	1100	50	17	BV	18
10 9 4dcd	Lundquist	110	1120	110	60	BV	30
10 9 5bcd	Williams	140	1110	137	92	BV	50
10 9 5cca	R. Hodges	100	1110	93		BV	30
10 9 5ccb	G. Hodges	134	1110	134+	80	BV	30
10 9 5ccc	L. Hodges	137	1125	137+	70	BV	30
10 9 5cdb	Woodard	125					
10 9 5ddc	L. Carver	84	1070	79	47	BV	30
10 9 6ada	C. Meyer	122	1130	119	60	BV	100
10 9 6bbb	S. Torrey	140	1090	140+	95	BV	40
10 9 7acc	J. Wilson	78	1050	72	40	BV	15
10 9 7acd	Anderson	69	1070	60	25	BV	30
10 9 7baa	D. Torrey	110	1060	102	65	BV	10
10 9 7ccc	R. Johnson	80	1040	43	25	BV	10
10 9 8aca	J. Wilson	78	1080	72	50	BV	15
10 9 8bbd	Hodges	187	1180	181		BV	20
10 9 8bbd	V. Claeys	70	1090	65	50	BV	30
10 9 8dac	Heideman	65	1040	40	20	BV	13
10 9 9cad	St. George	103	1030	98	79	BV	35
10 9 9bdc	D. Burgess	70	1090	16	19	BR	16
10 9 10baa	T. Vera	100	1150	70	52	BV	4
10 9 10bdd	Anderson	116					
10 9 10caa	J. Quinn	92	1150	9	65	BV	15
10 9 10caa	J. Fair	92	1150	92	65	BV	15
10 9 10bbc	A. Virgin	100	1120	68	50	BV	25
10 9 10ddd	Eschwein	140	1100	5	78	BR	4
10 9 11cca	M. Toy	110	1110	20		BV	8
10 9 11ccc	J. Estes	50	1090	38	20	BV	8
10 9 11dbb	Williams	53	1150	35	30	BR	1.5
10 9 12abb	M. Toy	85	1090	57	40	BV	18
10 9 14adc	H. Johnson	36	985	36+	15	KA	100
10 9 14dac	R. Shaw	54	990	54+	19	KA	100
10 9 14dbc	W. Taylor	69	985	69	15	KA	1200
10 9 14dbc	H. Johnson	36	980	36+	15	KA	100
10 9 14dbb	L. Johnson	50	980	50+	13	KA	150
10 9 15aca	P. Gardner	50	1050	28	22	BV	8
10 9 15abd	Drouillard	100	1100	77	58	Q	
10 9 15bbb	Oppenlander	60	1100	28	20	Q	5
10 10 1bda	Sylvestor	54	970	48	12	UA	1200
10 10 2cbc	Dr. Otto	64	980	64	24	KA	2000
10 10 2dcb	D. Crest	49	965	49+	18	KA	25
10 10 3bab	Weixelman	82	1500	82+	32	KA	1500
10 10 3ddd	Carnahan	83	980	82	16	KA	1560
10 10 4aaa	D. Doily	47	990	7	25	KA	50

Well location	well owner	well depth	surface elev.	depth to bedrock	depth to water	aquifer type	well yield (GPM)
10 10 5acc	Sackrider	83	1080	83+	65	BV	50
10 10 5bbb	K. Lamb	86	1090	79	57	BV	6
10 10 5bcd	Sackrider	94	1010	77	54	KA	12
10 10 6baa	Douglas	80	1120	23	20	BV	8
10 10 6bbb	Butler	120	1120	100	47	BV	3
10 10 6bbd	Huppe	100	1160	74		BV	15
10 10 6bdd	Hildreth	90	1165	78	60	BV	8
10 10 9bbb	Wamego #7	54	980	54+	30	KA	500
10 10 9bcc	Wamego #8	54	990	54+	32	KA	500
10 10 10aba	Fairchild	50	970	50+	30	KA	300
10 10 10abb	Fairchild	47	980	47+	25	KA	40
10 10 10bbb	Wamego,	79	980	79	32	KA	
10 10 11aad	Slagle	100	1030	48	45	KA	6
10 10 11baa	Carnahan	44	965	44+	27	KA	40
10 10 17bbb	Grothjan	50	990	28	19	KA	10
10 10 17caa	Doperalski	40	970	40+	15	KA	50
10 10 17ddc	Doperalski	60	980	60+	15	KA	100

APPENDIX   IX

Appendix IX. The calcite, dolomite, and gypsum saturation
indices and the ionic strengths of the sample waters.

Sample and Well Number	Owners Name	Aquifer Type	S.I. Calcite	S.I. Dolomite	S.I. Gypsum	Ionic Strength
8 9 35cdd#1	O. Blanka	BR	-.111	.045	-.79	.02555
8 9 35cdd#2	O. Blanka	BR	-.188	.047	-1.2	.01828
8 9 35cdd#3	O. Blanka	BR	-.248	.195	-1.1	.01786
9 8 26dbd#1	Thiele	BR	-.015	-.007	-1.54	.01179
9 8 26dbd#2	Thiele	BR	-.100	-.085	-1.8	.01186
9 8 26db#d3	Thiele	BR	-.249	-.220	-1.9	.00994
9 9 4ddc#1	Ebert	UA	.285	.415	-1.38	.01387
9 9 4ddc#2	Ebert	UA	.014	.349	-1.67	.01322
9 9 4ddc#3	Ebert	UA	-.092	.521	-1.7	.01302
9 9 4ddc#4	Ebert	UA	-.209	-.235	-1.75	.01358
9 9 11daa	Barth.	UA	-.095	-.513	0L	.01337
9 9 15dbc#1	Patnode	BV	.076	.172	-2.01	.00813
9 9 15dbc#2	Patnode	BV	-.107	-.011	-2.11	.00789
9 9 15dbc3#	Patnode	BV	-.307	-.360	-2.12	.00793
9 9 24add#1	Warnow	BV	-.821	-1.18	-2.69	.00472
9 9 24add#2	Warnow	BV	-1.25	-2.52	0L	.00386
9 9 24add#3	Warnow	BV	-1.08	-2.21	-3.33	.00405
9 9 24add#4	Warnow	BV	-1.79	-2.77	-2.75	.00386
9 9 30baa#1	O. Ubel	BV	-.377	-.341	-2.58	.00741
9 9 30baa#2	O. Ubel	BV	-.367	-.547	-3.15	.00560
9 9 30baa#3	O. Ubel	BV	-.785	-1.23	0L	.00588
9 9 30baa#4	O. Ubel	BV	-.305	-.397	-3.26	.00662
9 9 33adb#1	Piper	BV	-1.45	-2.19	-2.95	.00265
9 9 33adb#2	Piper	BV	-1.73	-3.24	0L	.00203
9 9 33adb#3	Piper	BV	-1.77	-3.43	-3.68	.00208
9 9 33adb#4	K. Piper	BV	-1.07	-2.03	0L	.00258
9 9 34abb	Lintz	BV	-.918	-1.98	-2.90	.00253
9 9 36bcc	Hudson	BV	-1.27	-2.52	-2.90	.00320
9 10 6aab	Brunkow	BR	-.194	-.014	-1.1	.01531
9 10 3cdc	L. Blanka	Q	-.003	.074	-1.3	.02595
9 10 14aba	Whearty	Q	.369	.945	-1.43	.01663
9 10 16cac#1	Umscheid	Q	-.425	-.592	-.95	.01984
9 10 16cac#2	Umscheid	Q	-.998	-1.63	-1.03	.01932
9 10 16cac#3	Umscheid	Q	-.755	-.925	-1.39	.01378
9 10 16cac#4	Umscheid	Q	-.105	.179	-1.20	.01752
9 10 28bda#1	Bammes	UA	-.703	-1.76	-2.5	.00544
9 10 28bda#2	Bammes	UA	-1.24	-2.48	-3.19	

KA = Kansas River alluvium   BV = buried valley sediments
BR = bedrock   Q = glacial sediments
UA = upland stream alluvium
                    Note:Ionic Strength is in Moles per Liter

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Sample and Well Number	Owners Name	Aquifer Type	S.I. Calcite	S.I. Dolomite	S.I. Gypsum	Ionic Strength
9 10 28bda#3	Bammes	UA	-1.36	-2.73	-3.15	.00433
9 10 32cbc#1	Seele	BV	-.010	-.007	-2.7	.00582
9 10 32cbc#2	Seele	BV	-.494	-1.00	-3.16	.00586
9 10 32cbc#3	Seele	BV	-.582	-3.16	-3.28	.00586
9 10 35bbc	Boyce	UA	-.062	-.017	-1.9	.01105
9 11 6bcc	Peterson	UA	.413	.959	-.57	.00228
10 8 2bcd	T. Ubel	BV	-.108	-.159	-1.36	.07188
10 8 12cab#1	Kern	KA	-.285	-.488	-2.1	.00914
10 8 12cab#2	Kern	KA	-.272	-.481	-2.44	.00941
10 8 12cab#3	Kern	KA	-.225	-.382	-2.26	.00809
10 9 3cdc#1	Altwegg	BR	-.008	.302	-2.23	.00
10 9 3cdc#2	Altwegg	BR	-.435	-.858	-2.66	.00607
10 9 3cdc#3	Altwegg	BR	-.386	-.284	-2.3	.00566
10 9 5ddc#1	Carver	BV	-1.27	-2.51	-2.41	.00397
10 9 5ddc#2	Carver	BV	-1.33	-3.40	DL	.00291
10 9 5ddc#3	Carver	BV	-1.17	-2.20	-2.40	.00470
10 9 5ddc#4	Carver	BV	-1.11	-2.28	-2.46	.00446
10 9 6ada	Meyer	BV	-.999	-2.00	-2.5	.00332
10 9 7baa	Torrey	BV	-2.39	-4.95	-2.61	.00183
10 9 14dac	Shaw	KA	.013	-.304	-2.0	.00939
10 9 15bbb#1	Oppen.	Q	.476	.805	.24	.04217
10 9 15bbb#2	Oppen.	Q	-.195	-.246	.01	.03321
10 9 15bbb#3	Oppen.	Q	.006	.202	-.29	.03537
10 9 15bbb#4	Oppen.	Q	-.148	-.065	-.35	.03355
10 10 1bda	Slyvestor	KA	.035	-.093	-2.06	.01005
10 10 6bdd#1	Hildreth	BV	-1.40	-2.61	-2.83	.00383
10 10 6bdd#2	Hildreth	BV	-1.30	-2.46	-2.84	.00364
10 10 6bdd#3	Hildreth	BV	-1.87	-3.50	-1.40	.00362
10 10 10abb#1	Fair.	KA	.314	.515	-.9	.01859
10 10 10abb#2	Fair.	KA	.089	.293	-1.1	.01554
10 10 10abb#3	Fair.	KA	-.221	-.283	-1.1	.01779
10 10 10abb#4	Fair.	KA	.204	.466	-1.15	.01681
10 10 17cac	Ooperalski	KA	.269	.224	-1.13	.01215

KA = Kansas River alluvium   BV = buried valley sediment
BR = bedrock   Q = glacial sediments
UA = upland stream alluvium
            Note:Ionic Strength is in Moles per Liter

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APPENDIX X

Appendix X. The computer printouts showing the results from the
analyses of the variance between the ions and saturation indices of
calcite, dolomite, and gypsum of the buried valley sediments and the
other hydrologic groups.

C1 = The Data for the Buried Valley Sediments

C6 = The Ion Data for the Other Hydrologic Units

C2 = The Saturation Indices of the Other Hydrologic Units

N = Number

F = Computed F Statistic

SS = Variance

MS = Mean Square

STDEV = Standard Deviation

```
MTB > RETRIEVE 'FE-DATA'
MTB > JOIN C2 C3 C4 C5, PUT INTO C6
MTB > AOVONEWAY C1,  C6

ANALYSIS OF VARIANCE
SOURCE    DF       SS        MS        F
FACTOR     1     19.82     19.82     5.08
ERROR     67    261.27      3.90
TOTAL     68    281.10
                                INDIVIDUAL 95 PCT CI'S FOR MEAN
                                BASED ON POOLED STDEV
LEVEL      N      MEAN     STDEV   ---------+---------+---------+-------
C1        29     0.094     0.176   (----------*----------)
C6        40     1.180     2.584                  (---------*-------)
                                   ---------+---------+---------+-------
POOLED STDEV =    1.975              0.00      0.75      1.50
MTB > SAVE 'FE-DATA'
MTB > RETRIEVE 'SR-DATA'
MTB > JOIN C2WC3 C4 C5,CBUT INTO C6
MTB > AOVONEWAY C1, C6
```

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```
ANALYSIS OF VARIANCE
SOURCE    DF      SS        MS       F
FACTOR    1     104.4     104.4    3.96
ERROR    68    1792.8      26.4
TOTAL    69    1897.2
                                   INDIVIDUAL 95 PCT CI'S FOR MEAN
                                   BASED ON POOLED STDEV
LEVEL     N     MEAN      STDEV    --------+---------+---------+-------
C1       30     0.194     0.135    (--------*--------)
C6       40     2.662     6.779                  (-------*-------)
                                   --------+---------+---------+-------
POOLED STDEV =   5.135               0.0      2.0      4.0
MTB > RETRIEVE 'MN-DATA'
MTB > JOIN C2 C3 C4 C5, PUT INTO C6
MTB > AOVONEWAY C1, C6

ANALYSIS OF VARIANCE
SOURCE    DF      SS        MS       F
FACTOR    1     0.337     0.337    2.39
ERROR    67     9.471     0.141
TOTAL    68     9.808
                                   INDIVIDUAL 95 PCT CI'S FOR MEAN
                                   BASED ON POOLED STDEV
LEVEL     N     MEAN      STDEV    -----+---------+---------+---------+
C1       31    0.0700    0.1910    (-----------*----------)
C6       38    0.2105    0.4758              (----------*----------)
                                   -----+---------+---------+---------+
POOLED STDEV =   0.3760             0.00     0.12     0.24     0.36
MTB > RETRIEVE 'CA-DATA'
MTB > AOVONEWAY C1, C6

ANALYSIS OF VARIANCE
SOURCE    DF      SS        MS       F
FACTOR    1    264191    264191   59.40
ERROR    68    302459      4448
TOTAL    69    566650
                                   INDIVIDUAL 95 PCT CI'S FOR MEAN
                                   BASED ON POOLED STDEV
LEVEL     N     MEAN      STDEV    ------+---------+---------+---------
C1       30    39.83     18.20     (----*----)
C6       40    163.98    86.66                          (---*---)
                                   --------+---------+---------+-------
POOLED STDEV =   66.69               50       100      150
MTB > RETRIEVE ;MG-DATA'
MTB > ÆOVONEWAY C1, C6

ANALYSIS OF VARIANCE
SOURCE    DF      SS        MS       F
FACTOR    1    3745.8    3745.8   48.59
ERROR    68    5242.2      77.1
TOTAL    69    8987.9
                                   INDIVIDUAL 95 PCT CI'S FOR MEAN
                                   BASED ON POOLED STDEV
LEVEL     N     MEAN      STDEV    ------+---------+---------+---------
C1       30    5.231     3.150     (-----*----)
C6       40    20.013    11.271                       (---*----)
                                   -------+---------+---------+--------
POOLED STDEV =   8.780               6.0      12.0     18.0
```
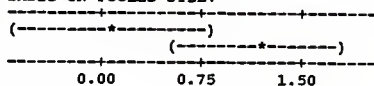
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```
MTB > RETRIEVE 'K-DATA'
MTB > AOVONEWAY C1, C6

ANALYSIS OF VARIANCE
SOURCE     DF       SS       MS       F
FACTOR      1    138.6    138.6    4.94
ERROR      68   1909.1     28.1
TOTAL      69   2047.7
                                    INDIVIDUAL 95 PCT CI'S FOR MEAN
                                    BASED ON POOLED STDEV
LEVEL      N     MEAN    STDEV    ----------+---------+---------+------
C1        30    2.057    0.640    (--------*----------)
C6        40    4.900    6.975                    (---------*-------)
                                   ----------+---------+---------+------
POOLED STDEV =    5.299                    2.0       4.0       6.0
MTB > RETRIEVE 'NA-DATA'
MTB > AOVONEWAY C1, C6

ANALYSIS OF VARIANCE
SOURCE     DF       SS       MS       F
FACTOR      1     4456     4456    7.54
ERROR      68    40196      591
TOTAL      69    44652
                                    INDIVIDUAL 95 PCT CI'S FOR MEAN
                                    BASED ON POOLED STDEV
LEVEL      N     MEAN    STDEV    ---------+---------+---------+-------
C1        30    20.63     7.54    (---------*-------)
C6        40    36.76    31.44                 (--------*------)
                                   ---------+---------+---------+-------
POOLED STDEV =    24.31                    20        30        40

MTB > RETRIEVE    -DATA'
MTB > AOVONEWAY C1 C6


ANALYSIS OF VARIANCE
SOURCE     DF       SS       MS       F
FACTOR      1   503279   503279   67.35
ERROR      68   508120     7472
TOTAL      69  1011399
                                    INDIVIDUAL 95 PCT CI'S FOR MEAN
                                    BASED ON POOLED STDEV
LEVEL      N     MEAN    STDEV    --+---------+---------+---------+----
C1        30   172.93    78.36    (---*---)
C6        40   344.27    91.99                         (---*---)
                                   --+---------+---------+---------+----
POOLED STDEV =    86.44          150       225       300       375
MTB > RETRIEVE 'SO4-DATA'
MTB > AOVONEWAY C1 C6

ANALYSIS OF VARIANCE
SOURCE     DF       SS       MS       F
FACTOR      1   306460   306460   11.84
ERROR      68  1759634    25877
TOTAL      69  2066093
```

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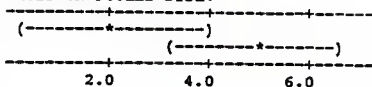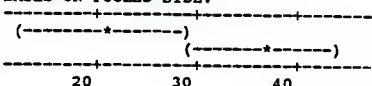
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                                         INDIVIDUAL 95 PCT CI'S FDR MEAN
                                         BASED DN PDOLED STDEV
LEVEL        N        MEAN     STDEV    --------+---------+---------+---------
C1          30        11.3      16.9    (--------*------)
C6          40       145.0     211.9                        (-----*------)
                                        --------+---------+---------+---------
POOLED STDEV =      160.9                    0        75        150
MTB > RETRIEVE 'CL-DATA'
MTB > ADVDNEWAY C1,C66

ANALYSIS DF VARIANCE
SDURCE       DF        SS        MS        F
FACTOR        1      7088      7088     17.70
ERROR        68     27227       400
TDTAL        69     34315
                                         INDIVIDUAL 95 PCT CI'S FDR MEAN
                                         BASED DN POOLED STDEV
LEVEL        N        MEAN     STDEV    --+----------+---------+---------+----
C1          30        5.82      3.67    (------*------)
C6          40       26.15     26.23                      (-----*-----)
                                        --+----------+---------+---------+----
POOLED STDEV =       20.01               0        10        20        30
MTB > RETRIEVE NND3-DATA'
MTB > AOVONEWAY C1 C6

ANALYSIS DF VARIANCE
SDURCE       DF        SS        MS        F
FACTDR        1      6426      6426      2.23
ERRDR        34     97921      2880
TOTAL        35    104347
                                         INDIVIDUAL 95 PCT CI'S FOR MEAN
                                         BASED ON POOLED STDEV
LEVEL        N        MEAN     STDEV    -----+---------+---------+---------+-
C1          16       16.56     23.12    (----------*----------)
C6          20       43.45     68.79                 (----------*---------)
                                        -----+---------+---------+---------+-
PODLED STDEV =       53.67               0        24        48        72
```
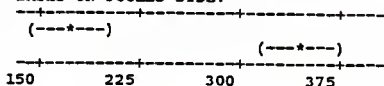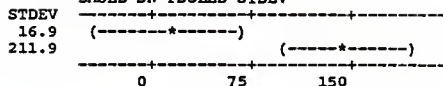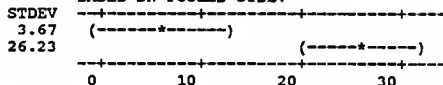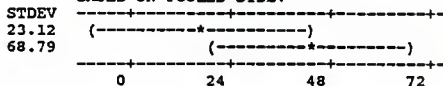
```
MTB > RETRIEVE ¦CALCITE-DATA'
MTB > AOVONEWAY C1 C2

ANALYSIS OF VARIANCE
SOURCE      OF        SS        MS         F
FACTOR       1      9.273     9.273     29.25
ERROR       68     21.560     0.317
TOTAL       69     30.833

                                    INOIVIOUAL 95 PCT CI'S FOR MEAN
                                    BASEO ON POOLEO STOEV
LEVEL       N       MEAN      STOEV   ----+---------+---------+---------+--
C1          30     -0.9065    0.7268  (------*-----)
C2          40     -0.1710    0.4001
                                                           (-----*----)
POOLED STOEV =     0.5631             ----+---------+---------+---------+--
MTB > RETRIEVE 'DOLO-DATA'                -0.96     -0.64     -0.32     -0.00
MTB > AOVONEWAY C1 C2

ANALYSIS OF VARIANCE
SOURCE      OF        SS        MS         F
FACTOR       1      49.35     49.35     46.00
ERROR       68     72.95      1.07
TOTAL       69    122.31

                                    INOIVIOUAL 95 PCT CI'S FOR MEAN
                                    BASEO ON POOLEO STOEV
LEVEL       N       MEAN      STOEV   -+---------+---------+---------+-----
C1          30     -1.901     1.288   (----*----)
C2          40     -0.204     0.798
                                                       (---*----)
POOLEO STOEV =     1.036              -+---------+---------+---------+-----
MTB > RETRIEEE 'GYPSUM-OATA'          -2.25     -1.50     -0.75     -0.00
MTB > AOVONEWAY C1 C2

ANALYSIS OF VARIANCE
SOURCE      OF        SS        MS         F
FACTOR       1      9.963     9.963     10.73
ERROR       67     62.239     0.929
TOTAL       68     72.202

                                    INOIVIDUAL 95 PCT CI'S FOR MEAN
                                    BASED ON POOLEO STOEV
LEVEL       N       MEAN      STDEV   ----+---------+---------+---------+--
C1          30     -2.2117    1.1270  (--------*-------)
C2          39     -1.4451    0.8177
                                                     (-------*-------)
POOLED STOEV =     0.9638             ----+---------+---------+---------+--
                                         -2.40     -2.00     -1.60     -1.20
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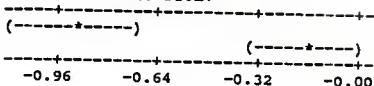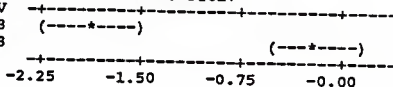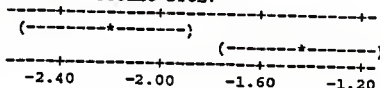
HYDROGEOLOGY OF A BURIED VALLEY AND ADJACENT
AQUIFERS IN SOUTHWEST POTTAWATOMIE COUNTY, KANSAS

by

JOSEPH GUSTAV RICHTER

B. A., College of Saint Thomas, 1982

AN ABSTRACT OF A THESIS

submitted in partial fulfillment of the

requirements for the degree

MASTER OF SCIENCE

Geology

KANSAS STATE UNIVERSITY
Manhattan, Kansas

1988

ABSTRACT

The hydrogeology of a buried, pre-glacial valley and the adjacent
aquifers in southwest Pottawatomie County were examined using the
data from water well logs and by analysing water samples from the
area.

The buried, pre-glacial valley is 12 miles long, 3 to 5 miles wide,
and the sediments filling the valley may produce up to 450 gallons of
water per minute. Seventy water samples were collected on four occa-
sions from the buried, pre-glacial valley and the surrounding geolo-
gic units. The water samples were analyzed for major and selected
trace constituents.

The mean ion concentrations of the water samples were: 94 mg/l
Ca; 30 mg/l Na; mg/l Mg; 3.7 mg/l K; mg/l HCO_3; 90 mg/l SO_4; 16 mg/l
Cl; and 31 mg/l NO_3. Water samples from all the hydrologic units,
except the buried, pre-glacial valley, were similar in composition.
Most waters were Ca–HCO_3 type water. Water samples from the hydro-
logic units surrounding the buried, pre-glacial valley had mean satu-
ration indices of calcite ranging from -.008 to -.305; mean saturation
indices of dolomite ranging from -.053 to   -.550; and mean satura-
tion indices of gypsum ranging from -.77 to -1.99. Water samples from
Kansas River alluvium contained large concentrations of potassium from
pollution by fertilizers. Water samples from the bedrock contained
high concentrations of strontium. The water samples with the highest
concentrations of calcium, magnesium, sodium, iron, manganese, chlo-
ride, sulfate and nitrate, and the largest saturation indices were
obtained from the glacial deposits. Water samples from the glacial

deposits are believed to contain large amounts of dissolved solids due to recharge from the bedrock.

The water samples with the lowest concentrations of dissolved constituents and the smallest saturation indices were obtained from the buried, pre-glacial valley. The mean ionic concentrations in water samples from the buried, pre-glacial valley were: 40 mg/l Ca; 21 mg/l Na; 5.2 mg/l Mg; 2.1 mg/l K; 179 mg/l HCO_3 ; 11 mg/l SO_4 ; 5.8 mg/l Cl ; and 23 mg/l NO_3 The mean saturation indices of the water samples obtained from the buried, pre-glacial valley were: calcite, -.98; dolomite, -1.90; and gypsum, -2.21. The low amounts of dissolved solids in water from the buried, pre-glacial valley are due, in part, to the buried, pre-glacial valley containing large amounts of insoluble sediments.

The quality of water from the buried, pre-glacial valley, bedrock, upland stream and Kansas River alluvium appears to change within three months of a major change in precipitation. The quality of water from the glacial deposits appears to change within six months of a major change in precipitation.

The small amounts of dissolved solids in water samples from the buried, pre-glacial valley suggest that the buried, pre-glacial valley receives little recharge from other hydrologic units, and that most of the water in the buried, pre-glacial valley originates from the infiltration of precipitation and surface water. The buried, pre-glacial valley recieves small amounts of recharge from a bedrock, topographic high located between Wamego and St. George, Kansas, and from glacial deposits to the west of the buried, pre-glacial valley.